Determination of Losing and Gaining Reaches in Arid and Semi-Arid Environments of NSW

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of the requirements for the degree of
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CERIFICATION OF ORIGINALITY

I hereby certify that the substance of the material used in this study has not been submitted already for any degree and is not currently being submitted for any other degree and that to the best of my knowledge any help received in preparing this thesis, and all reference material used, have been acknowledged.

Dawit Berhane

Brisbane, 28th October 2015

As part of this PhD study, hydrological processes of a riffle/pool junction were studied in a field environment.

Acknowledgments

Reaching the final stage of this PhD was a challenging exercise for numerous reasons. I suppose, the end of this study has been a relief to my family and me. In this journey, I would like to remind the reader that completion of this work would not have been achieved without a genuine technical and moral support of many colleagues and friends. Therefore, I acknowledge my warmest gratitude to the following people:

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Abstract

During the last decade, surface groundwater connectivity has become a major issue for water resources management in NSW. Consequently, as part of this PhD study, I have applied heat an environmental tracer to study infiltration/exfiltration in two contrasting hydro-geomorphological environments in NSW, Australia (Peel, Cockburn and Gunnedah sites in the Namoi Catchment, and Baldry and Sloans sites in the Central West Catchment). In the context of this thesis, the hydrological processes targeted of relevance to water resources management are grouped into three main issues.

Assessment of infiltration is a prerequisite for environmental flow management. Thus, the first issue focused on the determination of infiltration/exfiltration at a reach scale in intermittent and ephemeral streams in NSW. In the case of the Cockburn Valley (head waters of Peel River and an unregulated stream), data collection extended for more than 4 years and two ‘extreme’ hydrologic events that occurred in late 2008 were captured. The maximum recorded stream stages were 6.1 and 3.2 m, with a corresponding stream leakage of 4×10^{-4} and 2.8×10^{-4} m/s, respectively.

The instrumentation in the Central West Catchments (Baldry and Sloans) was deployed for a shorter period. Nevertheless, the results from model simulations provide an opportunity to understand the nature of surface–groundwater connectivity issues of a typical ephemeral stream. Hydrologic data suggests that the predominant part of the thermographs displays a conductive heat transport mechanism. However, rare storm events during the measurement

period were captured by an abrupt change in streambed temperature. Infiltration rates varied from 0 to 9.0×10^{-6} m/s, during the instrumentation period.

The second issue I assessed was the dynamic nature of streambed hydraulic conductivity, which was inferred from the streambed thermal data for the Kootingal site. Scouring of the streambed sediments is likely to occur after flood events, while clogging takes place during low flow events. In the Cockburn Valley, clogging and scouring processes may have been enhanced by human induced changes.

The last issue I explored was the relevance of scale on surface-groundwater connectivity study. Prior to the field experimentation, the Cockburn Valley was considered as a groundwater dominated catchment. However, analysis of streambed temperature in conjunction with hydrologic data from different reaches of the study area suggests that stream loses at a reach and sub-catchment scales. It appears that small-scale exchanges that take place at a riffle/pool junction do not provide enough hydrologic evidence for groundwater dominance in the study area. Nevertheless, the small-scale processes are of relevance to stream health. Meyer (1977) describes a healthy stream as having a "…sustainable and resilient ecological structure over time while continuing to meet societal needs and expectations."

Thus, this thesis provides a case study, using heat tracing as a tool to enhance hydro(geo)logic knowledge on surface groundwater connectivity in selected catchments in NSW; it clearly demonstrates the presence of surface water groundwater exchanges in the physical environments of NSW at different spatial and time scales. The recharge/discharge estimates using this method can be further be up-scaled when used in conjunction with other tracers and hydrologic information.

Chapter 1 Motivation and objectives

"For in the end we will conserve only what we love;
we will love only what we understand;
and we will understand only what we are taught."

Baba Dioum, 1968

Preface

A growing human population, expanding urbanisation and industrialisation, and increasing demands for food production are placing more pressure on the world's groundwater supplies (Famiglietti et al., 2011; Leblanc, 2009; Rosegrant et al., 2009; Sophocleous, 2000; Voss et al., 2013). More recently, uncertainties in rainfall distributions in space and time, associated with climate variability, and expansion of mining related activities in different parts of the world are having further impacts on the fragile water resources sector of arid and semi-arid regions of the world.

Groundwater is the largest readily available source of fresh water on our planet (lakes and reservoirs 0.13×10^6 km^3, river channels < 0.01×10^6 km^3, groundwater 60×10^6 km^3 (Freeze and Cherry, 1979) and the basis for life and socio-economic development in many areas of the world. Given the critical role of water for our survival and social and economic development, it is difficult to comprehend why inappropriate sector strategies are formulated and implemented, related to water resources management, even when they could have been realised to be problematic based on available knowledge at the time (Ronen et al., 2011).

Finding a right balance between socio-economic motivated decisions and technical advice is a challenge in water resources management. Unfortunately, water resources plans based solely on socio-economic factors may not take into account long term impacts, which may occur beyond water planning time frames. The prevailing drought condition in California is a case study, which illustrates the lack of groundwater management may have undesired consequence on a long term of water resources management of an area. In July 2015, urban water use in California was cut by 31.3 percent during July[1].

In addition, although technical knowledge and recommendations on water resources research should be based on objective facts of the day, the author during the implementation phase of this project had to overcome side issues not linked to scientific endeavour. Some of these issues may be related to specific set of cultural values. For example, Mitchell (1994) argues that landscape should not be considered as an object to be seen?, "but as a process by which social and subjective identities are formed"; landscape, he writes, must be understood as a physical and "multisensory" medium "in which cultural meanings and values are encoded, whether they are put there by the physical transformation of a place in landscape gardening and architecture, or found in a place formed, as we say, 'by nature'" (Mitchell, 1994). Unfortunately, it is beyond the scope of this thesis to discuss aspects of culture in relation to hydro(geo)logic issues addressed in this thesis.

Putting aside ‘cultural values’ in the context of this thesis, accurately assessing and managing available and renewable water resources is more difficult in semi-arid regions, compared with water-rich areas, since the science base is limited, data are scarce and the humid zone experience is inappropriate applied to other cases (UNSECO, 2011). From a water resources

[1] http://www.huffingtonpost.com/entry/california-water-use_55df526fe4b0e7117ba92c9f?utm_hp_ref=california-drought

management perspective, some of the 'distinct' hydrological characteristics of arid and semi-arid regions are highlighted in the following paragraphs.

Evapotranspiration is generally the largest component of transmission losses in arid and semi-arid regions of the world. Channel transmission losses (TL) take place in (i) allogenic rivers, which are sourced almost entirely from upstream humid areas (e.g., the River Nile in Northern Sudan and Egypt) and commonly sustain perennial flow partly infiltrating in the alluvial system along the allogenic river, and (ii) endogenic rivers, which are sourced almost entirely within dryland environments and usually show an ephemeral (non–base flow) or intermittent flow (Bull and Kirkby, 2002; Lange, 2005). In Australia, most of the regional water resources models don't take into account TL explicitly.

From a surface-groundwater connectivity point of view, recharge/discharge processes in semi-arid/ephemeral river systems are distinctly different from the hydrology of humid/sub-humid environments. For example, in humid climates the regional aquifer intercepts the land surface and perennial streams are common. In contrast, in arid and semi-arid regions, the regional water table is usually disconnected or intermittently connected to the stream body (Sen, 2008). Due to the presence of a thick overburden or regolith layer and high evaporation rates, the regional water table is deep. In some cases, a perched/shallow groundwater system may exist above the regional/local groundwater systems (Cendón et al., 2010).

In arid to semi-arid climates, stream leakage in intermittent and ephemeral streams may provide localised groundwater recharge to the underlying regional aquifers (Izbicki et al., 2002; Stonestorm and Constantz, 2003). Present-day recharge tends to be narrowly focused in time and space (Stonestrom and Harrill, 2007). The literature review on recharge mechanisms of semiarid regions in this thesis indicates that distributed recharge is insignificant and only occurs when precipitation exceeds evaporation (Chapter 2 of this thesis).

Although there is no consensus in the inland catchments of NSW and the Northern MDB catchments, stream induced recharge is considered to be a major component of recharge (Berhane, 2006; Brownbil et al., 2011). Thus, information on the magnitude of stream induced recharge in ephemeral and intermittent streams would facilitate more effective water resources and ecosystem management (SKM, 2011). In order to increase the level of certainty, infiltration (maximum potential recharge) can be estimated using multiple methods (physical, piezometery, thermometry, seepage meters and channel water balance and stable isotopic techniques). However, some of the methods can be very expensive to apply (Healy, 2010).

1.1 Research objectives

The main objective of this research is to assess the potential application of heat as an environmental tracer to estimate infiltration/exfiltration rates and hydraulic conductivities on different hydrogeomorphological environments in New South Wales (NSW). Two focus catchments (Peel Valley near Gunnedah in the Namoi Catchment and Baldry in the Central West Catchment) in NSW were chosen. These catchments provided a combination of high stream flow variability and abundance of heat fluxes. This combination provides an ideal thermal field for the application of heat as an environmental tracer at different spatial and time scales and geological complexities. In addition, this research work is expected to provide a better understanding of recharge/discharge processes in the study area. The specific objectives of the PhD research work are to:

- Review the state of knowledge on surface groundwater connectivity in semi-arid and arid regions of NSW with a specific focus on the Namoi Valley;
- Visualise groundwater mounding in connected and disconnected groundwater systems using a physical based model and comprehend some of the underlying hydrological processes;
- Develop a relationship between water table and stream seepage rate for a given simple stream channel geometry and other hydraulic parameters;
- Demonstrate the potential application of the thermal method in ephemeral and intermittent in both unregulated and unregulated streams in NSW;
- Infer the dynamic nature of streambed hydraulic conductivity using thermal data;
- Delineate losing and gaining reaches using temperature as a tracer (Figure 1-1);
- Explore recharge/discharge processes at different spatial/time scales.

1.2 Outline of the thesis

This thesis is structured in six chapters. These chapters have a common link of surface-groundwater connectivity issues in arid and semi-arid regions of NSW. However, each chapter addresses a specific issue on surface groundwater connectivity and can be considered to be ‘independent’.

In Chapter 1, the scientific and the social relevance of this study is presented. The research motivation and objectives in this study are also stated in this chapter.

Chapter 2 provides a review of literature on transmission losses (TL) in semi-arid environments of Australia, with the main focus on the Namoi Valley. The primary objective of this work was to further understand the nature of the surface and groundwater interactions in the Namoi valley and identify knowledge gaps for further investigation. Here, several case studies from both regulated and unregulated sub-catchments in the Namoi Valley have been reviewed. Based on the review, heat as an environmental tracer was identified to be the most cost effective method for this research.

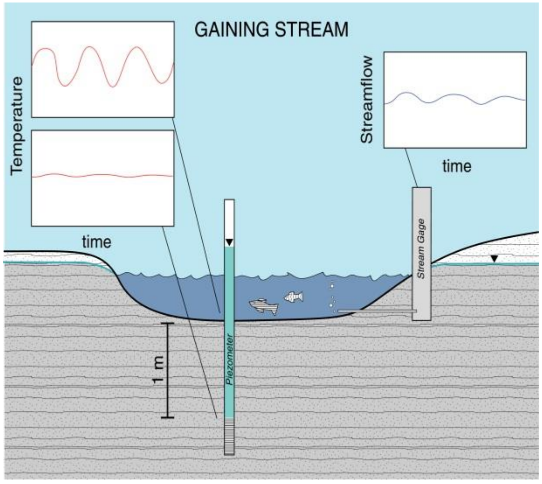
Chapter 3 visualises disconnected river systems in the Northern MDB. The chapter provides background information into our understanding recharge/discharge processes in the region. Prior to the application of heat as an environmental tracer, an insight into hydrological processes of ephemeral and intermittent streams was gained by simulating hypothetical head distributions using the VS2DT package.

Chapter 4 investigates the use of heat as a hydrological tracer for assessment of infiltration and near streambed water exchanges at two focus catchments in NSW. Among the environmental tracers used in hydrology, heat is considered to be a robust, inexpensive and effective tracer to assess near streambed water exchanges. This tracer was used in two contrasting hydro-geomorphological environments in NSW: Peel Valley near Gunnedah in the Namoi Catchment; Baldry in the Central West Catchment.

In ephemeral and intermittent streams, quantification of exchange of water between the stream and groundwater is problematic. In most of the cases the creek beds are dry and disconnected from the water-bearing formations (Figure 1-1). In this type of environment, surface groundwater interactions can assume even greater importance and affect the occurrence of flow (rather than the amount), the existence of static pools and the amount of water stored in the alluvial material beneath and adjacent to the channel (Hughes, 2010).

Temperature methods enable inference of streamflow timing on the basis of the combined transport of heat and fluid within the bed sediments (Constantz, 2001; Constantz and Thomas, 1996; Ronan et al., 1998). Thermal responses from within a streambed can be divided into the case when streamflow is absent (primarily conductive heat transport through the sediments) and the case when streamflow is present (primarily advective heat transport through the sediments (Blasch et al., 2004).

(a) Gaining Reach (b) Losing Reach



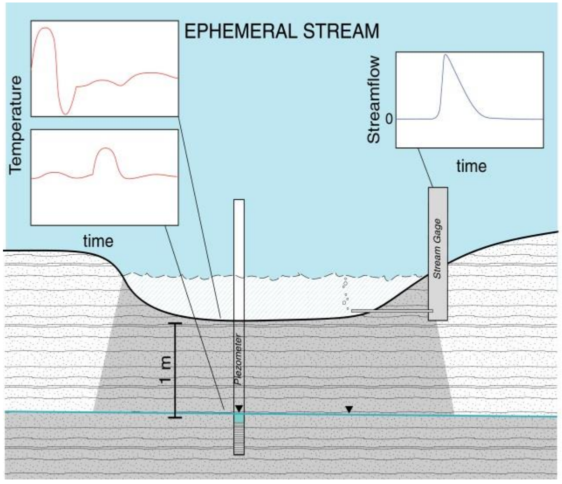


Figure 1-1. Groundwater temperatures generally are more stable than surface water temperatures. Therefore, gaining reaches are characterized by relatively stable sediment temperatures compared to stream-water temperatures. Conversely, losing reaches are characterized by more variable sediment temperatures caused by the temperature of the inflowing surface water (Constantz and Stonestrom, 2003b).

Chapter 5 demonstrates the use of long-term thermograph analysis for assessing the dynamic nature of streambed conductance in an intermittent stream, Cockburn River, NSW. In this chapter a segment thermograph is analysed, which represents the two flood events that occurred in late 2008. This illustrates the hydrologic regime prior to the flood events, the effect of floods on the temporal development of de-clogging process and low flow event.

Available hydrological information suggests that low-flow hydrologic regimes create a favourable environment for clogging of coarse streambed sediments and disintegration or scouring takes place during high and flood events.

Surface groundwater connectivity issues in the Cockburn Valley are complicated by human-induced changes superimposed on natural processes. The anthropogenic impacts on surface groundwater connectivity issues are caused by gravel mining of the past. Thus, chapter 4 summarises the current understanding of gravel extraction impacts on the physical environments of the Cockburn Valley (fluvial geomorphology, lowering of water table and threats on physical infrastructure – roads and bridges). This is considered to be a good example of a positive/negative feedback mechanism, which most deterministic models fail to take into account.

Chapter 6 provides a framework for discussion on the issue of scale in eco-hydrological science. One of the major challenges facing surface hydrology is the discrepancy between the models and observations against which validation occurs. Discrimination between overlapping space-time scales of the different processes and/or measurements is a daunting challenge in hydro-ecological sciences (Beven, 2001; Blöschl, 2001; Gentine et al., 2012).

For water allocation and planning purposes, information on water balance components is required at catchment and sub-catchment scales. In contrast, due to constraints in resources, time and incomplete knowledge of hydro-ecological systems, most of the hydro-ecological investigations are implemented at a reach or streambed scale. Thus, harmonising hydrological information obtained at different spatio-temporal scales is a challenging hydro-ecological issue.

As part of surface groundwater connectivity study in the Cockburn Valley, heat as an environmental tracer was used to get insight into hydrological processes that take place at different time/spatial scales. Temperature measured in both in-channel and off-channel piezometers have been used to better understand surface and groundwater connectivity issues in the study site. Within the scope of this study three 'distinct' spatial/time scales have been identified: Sub-catchment, reach and streambed scale.

More recently, analytical models (Hatch et al., 2010; Keery et al., 2007; Schmidt et al., 2006), have been used for delineation into gaining/losing reaches and estimation of fluxes at the streambed scale. Among these methods, the Bredehoeft method (BP) is of interest in the context of this study (Bredehoeft and Papadopulos, 1965b). It was originally applied for estimation of fluxes at a basin scale, where one of the main requirements is the assumption of a steady state condition.

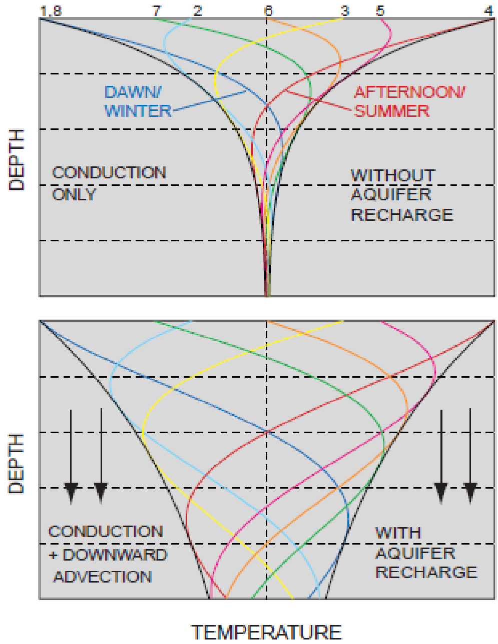


Figure 1-2. Recharging water advects diurnal and seasonal temperature fluctuations deeper into the profile than conduction alone. Numbered lines show successive temperature profiles over one daily or annual cycle (profiles 1 and 8 are identical, but separated by one cycle). The amount of downward shift in the bounding envelope depends on the rate of aquifer recharge (Nimmo, 2005).

From heat flux point of view, Parsons (1970) compartmentalized the subsurface into two zones: a surface zone, where temperatures fluctuate over time in response to energy exchange at the surface and a deeper geothermal zone that is largely isolated from the influence of climate (Healy, 2010). Within the geothermal zone, heat flows at a relatively steady state toward the surface as indicated by a vertical profile of increasing temperatures with depth, typically at rates of 0.01 to 0.02 ^{o}C/m (Healy, 2010). The depth to which the surficial zone extends varies. Short-term energy exchanges at land surface result in measurable temperature changes to depth of 0.1 to 0.3 m on daily basis and to about 10 m on annual basis (Anderson, 2005). Based on bore temperature profiles measured at the two sites in the Cockburn Valley, the depth of penetration of temperature extended up to 15m.

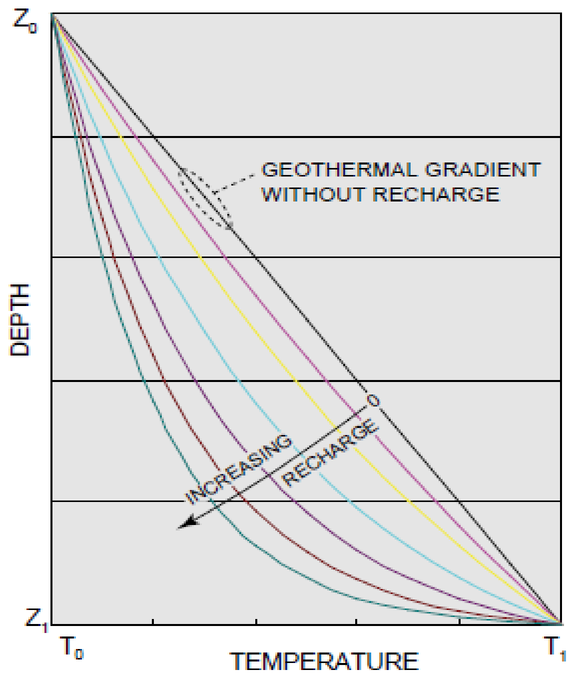


Figure 1-3. Recharging water in the deep unsaturated zone perturbs the steady geothermal profile produced by conduction alone. The degree of departure from the purely conductive profile indicates the amount of recharge. T0 and TL are temperatures at depths, Z0 and ZL, beneath the maximum penetration of seasonal fluctuations (Nimmo et al., 2005).

A further challenge to accurately represent the surface hydrologic state involves the scaling of physical surface hydrologic processes (Entekhabi et al., 1999). At present, our understanding of the scaling (both up and down) of such processes remains relatively unsophisticated (Gentine et al., 2012).

In Chapter 7, the final chapter, the main results of this thesis are summarised. Some conclusions and recommendations for future studies are provided.

Chapter 2 Transmission Losses in Arid and Semi-Arid Environments of Australia: A Literature Review

"It isn't the drought that is the problem. It is our delusions."

John Williams, 2003

Abstract

This chapter provides a review of literature on transmission losses (TL) of Australian dryland rivers, which are among the most hydrologically variable fluvial systems in the world. After reviewing the international and Australian literature, reports on TL in Australia are biased toward the more arid interior of the Lake Eyre Basin, where high TL and low water availability has created the most interest over the past three decades. In this arid region of the Australian continent, TL can reach up to 100% of the total runoff. However, with the introduction of water trade and environmental water recovery through buyback of water licences, the issue of surface groundwater connectivity and assessment of TL in stressed water systems has gained some momentum. Available published reports for a different region, the Murray Darling Basin (MDB), suggest that TL can vary from 20 to 100% in this part of Australia. In fact, TL occurs not only in the inland river systems but also on the coastal river systems, where river channels are predominantly intermittent or 'perennial'. The hydrologic regime of the eastern coastal river systems has been inferred by a regional average maximum no-flow spell duration developed for Queensland catchments.

As part of this study, several methods used for estimation of TL have been reviewed. The choice of a particular method for estimation of TL and recharge depends on the scale of investigation, the availability of resources and subjective preference of the user. Groundwater recharge, which cannot be measured directly, constitutes the smallest component of TL in

arid and semi-arid environments. It has been further found that indirect estimation methods introduce large uncertainty.

2.1 Introduction

Arid and semi-arid regions of the world are collectively termed drylands. They cover approximately 50% of the world's land area and support nearly 30% of its population (Figure 2-1). Hydrologic, geomorphic, and flow variability are some of the common characteristics of dryland rivers.

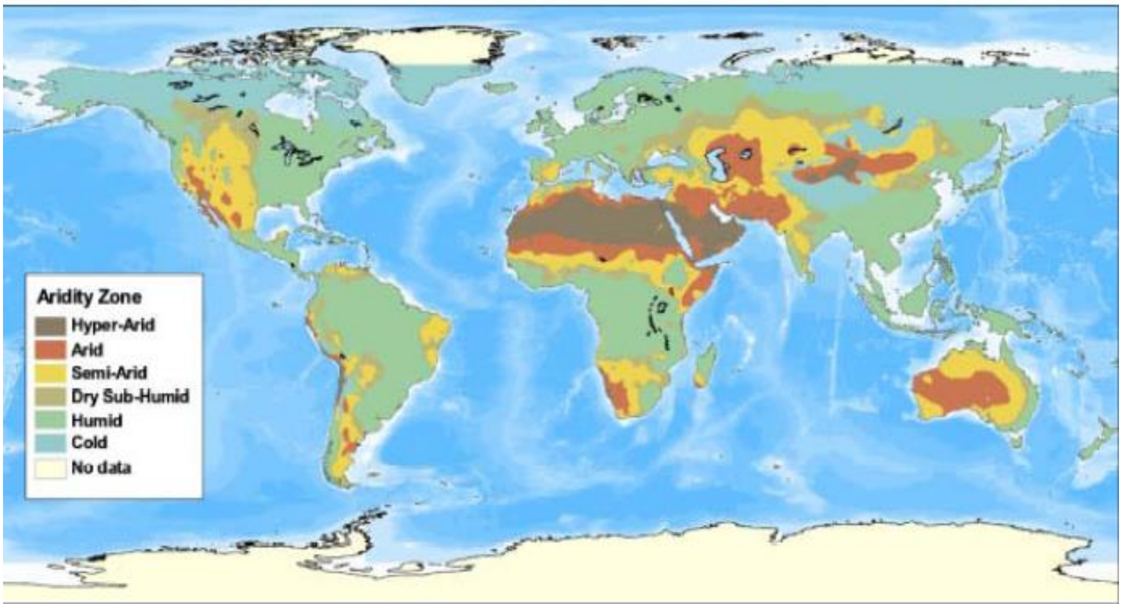


Figure 2-1. Map showing arid and semi-arid regions of the world. More than 90% of the Australia continent has been classified as arid and semi-arid region and 10% of eastern coastal regions are classified as humid (Source: UNEP/GRID, 1991; Accessed on 4/11/2012: knowledgebase.terrafrica.org/fileadmin/user.../drylands_WRI.pdf)

Hydrologically, dryland rivers can be grouped into allogeneic and endogenic river systems. In allogeneic river systems the main sources of water originate outside dryland regions, while endogenic systems source most of their flow from dryland regions (Lerner et al., 1990b; Nanson et al., 2002). In addition, allogenic rivers commonly sustain 'perennial' flow (e.g., the Nile River and the Euphrates River) whereas endogenic dryland rivers are usually intermittent or ephemeral with little or no baseflow and high channel transmission losses (Young and Kingsford, 2006). These hydrological differences have implications for water resources management, including issues of environmental flow, which is currently a complex management issue in Australia.

In dryland regions rainfall occurs as localised storms and shows high variability in space and time (Stafford-Smith and Morton, 1990). Evapotranspiration constitutes at least 95% of rainfall and the hydrologic balance is in a delicate state (Pilgrim, 1988). A prolonged wet or dry spell may change the whole nature of the hydrological characteristics. There is no reliable relationship between seasonal or annual runoff to rainfall and the magnitudes of the flood and the drainage basin areas are not directly related, except in small basins (Sharma, 1992). Ratios of mean annual precipitation to mean annual runoff for dry countries such as Australia and southern Africa are 9.8% and 8.6% respectively, and much lower than the world average of 48%.

Rivers in the semi-arid and arid regions are characterised by extreme flow variability, with distinct periods of runoff separated by periods of no flow (Knighton and Nanson, 2001). Based on multivariate analysis of the 52 largest river systems in the world (Puckridge et al., 1988) showed that Australian dryland rivers are considered to be the most hydrologically variable systems. The average coefficient of variation for annual runoff for dryland regions is 0.99 - much higher than 0.3 for the humid regions of North America, 0.2 for Europe and Asia (McMahon and Finlayson, 2003). For example, Cooper Creek, one of the major rivers in the Lake Eyre Basin, has the most variable hydrological regime in the world, followed by the Diamantina, Burdekin, Limpopo, Fitzroy, Vaal and Darling Rivers. In comparison, the

Colorado, Mississippi and Mekong rivers are far more predictable in their flows (Puckridge et al., 1988).

In Australia, 92% of the inland river systems are allogeneic (Thomas and Sheldon, 2000). These rivers originate in the relatively watered upland areas, but drain for most of their river course through the arid and semi-arid interior that yields little or no additional stream discharge (Parsons et al., 2009). Due to high rate of evapotranspiration, infiltration and filling in of waterholes and billabongs, the inland river systems experience large-scale losses. This natural loss of water, which is an inherent characteristic of dryland river systems, is called transmission losses (TL). It is envisaged that TL may occur in any climate type, but it is most common in arid and semi-arid regions.

Transmission losses occur as part of the hydrological cycle and should not have a negative connotation as wastage of water. In fact, TL should be recognised as a natural process of the hydrological cycle, where water, energy and momentum move from one compartment to another. In addition, TL support ecosystems and recharge local aquifers and regional groundwater (Renard, 1970).

At a basin scale, Abbott raised the issue of TL eloquently in his presentation to the Royal Society of NSW as early as 1880. Based on a simple observation on the components of the water balance equation in the Darling Basin, Abbott postulated: "I believe the area of the watershed of which the water passes Bourke, […] is about 140,000 square miles. The average rainfall at and about Bourke would be 16 inches [...]. Estimating [...] the rainfall throughout at 16 inches, and reducing over this large area in the usual way for evaporation and soakage, and in order to avoid the possibility of error reducing what remains by one-half, the river at

Bourke should be 200 feet deep and 200 yards wide, and flow all year round! At Bourke the river is sometimes nearly dry. Where then does all this water go? (J and Proceedings of the Royal Society of NSW, 1880, page 3)."

This review on TL was motivated by four main reasons. Firstly, seepage from losing streams or alluvial fans is considered to be the main sources of recharge in the inland rivers of NSW (Berhane, 2006; Brownbil et al., 2011). Secondly, estimating the impact of groundwater pumping on stream flows is a prerequisite for effective water resource and ecosystem management (Evans, 2007; MDBA, 2012). Thirdly, the recent introduction of water trading and release of water to the environment in Australia requires a better understanding of the components of TL (evapotranspiration, seepages and deep drainage), as this would affect volumes surface water availability. Fourthly, TL is suggested to have a vital role for maintenance of groundwater ecosystems (Doody et al., 2009; Holland, 2008).

The main objective of this review is to summarise the current knowledge on the spatial and temporal variation in transmission losses in ephemeral and intermittent streams in Australia, with particular focus on the Namoi Valley (NSW). The focus on Australia is also globally relevant. Australia, being the driest continent with the most variable riverflow in the world, can serve as a model for other dryland areas. Being a continent, it also encompasses a wide variety of TL. The four objectives of this review are:

- Review hydrological components of TL in arid and semi-arid regions;
- Review recharge and discharge mechanisms in waterholes and wetlands of arid Australia as a component of TL;

- Present commonly used methods for estimation of TL with emphasis on thermal methods for estimating recharge and discharge in ephemeral and intermittent river systems;
- Discuss management implications and gaps of knowledge in relation to TL in ephemeral and intermittent river systems. More importantly, which methods are cost effective for estimation TL at different spatial and temporal scales. One of the challenges in estimating TL is the need to integrate field based measurements carried out different spatial and temporal scales (Chapter 5 of this thesis).

More specifically, this review intends to invoke debate on the complex issue that is surface water and groundwater connectivity in semi-arid environments of NSW, and the Namoi valley in particular, where groundwater contributions to surface water bodies are considered to be negligible or non-existent.

2.1.1 Basic Concepts and Terminology

Ephemeral streams show high stream discharge variability and downstream decreases in discharge (Goodrich et al., 2004; Hughes and Sami, 1992; Hughes, 1992; Walters, 1990) as a result of storms covering only a portion of a watershed or due to the formation of series of billabongs or water holes in arid parts of Australia and Africa. In addition, substantial transmission losses result from high rates of evapotranspiration and infiltration into dry, unconsolidated alluvial beds, creating a positive feedback (Bull and Kirkby, 2002; Constantz et al., 1994; Goodrich et al., 2004; Renard, 1970; Tooth, 1999a).

Transmission losses are considered to be the most complex part the hydrological cycle, forming linkages between different compartments (surface waters bodies, unsaturated zone and groundwater systems). In a catchment context, landscape hydrologic connectivity refers to the maintenance of natural hydraulic connections of surface and subsurface flow between source, headwater, or contributing areas and downstream/down gradient receiving waters. (Nadeau and Rains, 2007) defined it as “the hydrologically mediated transfer of mass, momentum, energy, or organisms within or between compartments of the hydrologic cycle.” In arid land streams, this hydrologic connection occurs episodically during flood pulses, yet still provides a substantial amount of the mass, momentum, energy and organisms delivered to downstream ‘perennial’ waters, as well as to groundwater recharge (Levick et al., 2008).

Groundwater and surface water are intimately connected systems (Sophocleous, 2002; Sophocleous and Perkins, 2000; Winter, 1998; Woessner, 2000). Therefore, Sophocleous (2002) advocated a holistic approach to the management of water resources. He also stated that groundwater is a key component of environmental flows (EFs) and called upon hydrogeologists to participate in, and contribute to, debates about maintaining healthy riverine ecosystems. Most of the cited technical examples by Sophocleous (2002) are based on case studies from humid environments where the baseflow index (BFI) was used to illustrate the connectivity between surface and groundwater systems. The applicability of the BFI methodology for assessments of EFs in arid and semi-arid environments could be questioned however the general policy issues on EFs, he raised and promoted, do have relevance in all types of climatic environments.

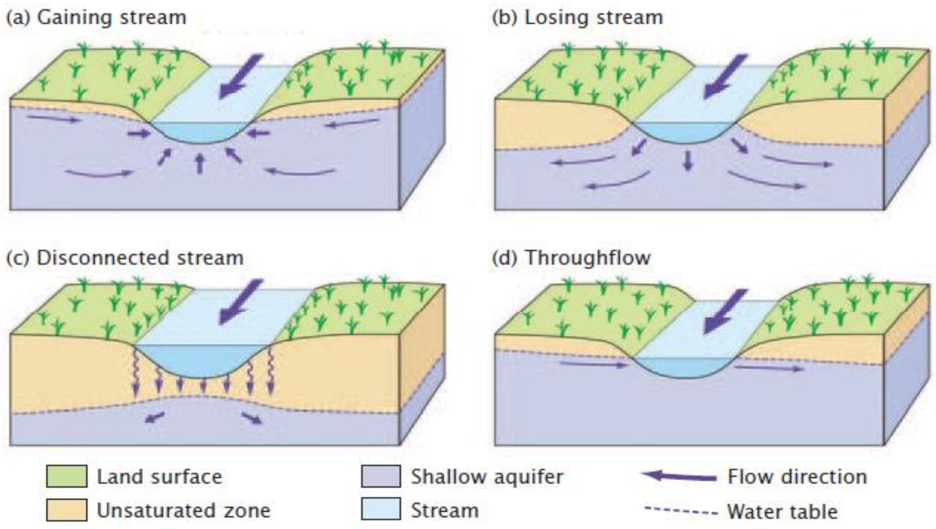


Figure 2-2. Types of stream-aquifer interactions: (a) connected gaining stream; (b) losing- connected stream; (c) disconnected stream; (d) throughflow system (adapted from Winter et al. 1998). During floods the stream stage rises, the river recharges the aquifer. The flow direction may reverse when the stage recedes. In addition, after floods the clogging layer is temporarily removed to facilitate connection between a river and groundwater system (Winter et al; 1998)

Providing water for the environment is more than just allocating water for the maintenance of river flows. Provisioning water for the environment requires allocations of water to maintain terrestrial, riparian, wetland, and stygian (i.e., groundwater–inhabiting organisms) ecosystems, which also require ground water for their survival (Murray et al., 2003).

Groundwater–dependent ecosystems (GDEs) are plant and animal communities which depend partly or completely on groundwater to maintain their current composition and functioning. Based on the definitions of (Eamus et al., 2006; SKM, 2001) GDEs include the following:

- Terrestrial vegetation that relies seasonally or episodically on ground water;
- River base flow systems, including aquatic, hyporheic, and riparian ecosystems that depend on ground water input, especially during dry periods;
- Aquifer and cave ecosystems, often containing diverse and unique fauna;

- Wetlands dependent on ground water influx all or part of the time; and
- Estuarine and nearshore marine ecosystems that rely on ground water discharge.

Streams, rivers, springs, and wetlands are all important GDEs that underscore the importance of surface water-ground water interactions.

Although TL is recognised as an important parameter of the water balance equation in arid and semi-arid environments, our knowledge still remains patchy and incomplete (NWC, 2012a). In recent research carried out in targeted catchments within the Murray Darling Basin (MDB), TL have been demonstrated to take place in four distinct hydrological environments (Brownbil et al., 2011; Brunner et al., 2011; Lamontagne et al., 2012): Gaining stream receiving water from local, intermediate or regional groundwater systems (Figure 2-2a); Losing connected to water table (Figure 2-2b); Disconnected groundwater system (Figure 2-2c), with water table sufficiently deep that the capillary fringe is below the base of stream sediments; Flow-through stream (Figure 2-2d).

2.2 Types of Transmission Losses

In Australia most of the investigations on TL were carried out in arid regions (Thomas and Sheldon, 2000). However, more recently with the introduction of water trading in the Murray Darling Basin, the issue of TL has gained some momentum and our knowledge on TL is evolving. Transmission losses along water supply routes are considered to be of four different types:

- Storage in waterholes/billabongs (ephemeral pools), channels and other wetlands;
- Evapotranspiration from the water surface;
- Seepage through the bottom of the river or channel;

- Leakage through the banks of the river or channel, and as overbank losses.

In addition, Inter-valley trading leads to changes in the location where water is extracted for use. This can affect transmission losses both along the major water supply routes and within irrigation areas. Changes in transmission losses will, in turn, have hydrological impacts (NWC, 2012b). For clarity and to minimise ambiguity, the four main types of transmission losses are discussed in more detail below, with the focus on the hydrological processes taking place in each component of TL.

2.3 Waterholes or Billabong Systems

Waterholes and flood-outs are two distinct geomorphic characteristics, which are common in dryland river systems of Australia and South Africa (Nanson et al., 2002). Waterholes are self-scouring sections of the channel or floodplain that may hold water between runoff events or during extended dry spells (Cendón et al., 2010; Knighton and Nanson, 1994b; Nanson et al., 2002). In the Australian dryland context, waterholes can be found in various fluvial environments[2] (Tooth, 1999b). They have been reported: (1) where ephemeral rivers traverse narrow bedrock gorges (Argue and Salte, 1977); (2) in the bed of otherwise dry, alluvial, ephemeral channels (Argue and Salte, 1977; Tooth, 1997); (3) at the junction of ephemeral tributaries with larger trunk channels (Tooth, 1997); (4) as part of extensive networks of intermittent anastomosing channels (Knighton and Nanson, 1994b); (5) on extensive muddy floodplains, away from primary anastomosing channels (Knighton and Nanson, 1994b); (6) on extensive, unchannelled, alluvial plains termed ‘floodouts’ (Tooth, 1999b). Some

[2] http://homepages.abdn.ac.uk/c.p.north/pages/DrylandRivers, accessed on 25/02/2012

waterholes are probably groundwater fed, while others retain water following local rainfall or flooding owing to sealing of the waterhole margins by clays dropping out of suspension.

Waterholes are common geomorphological features in the Channel Country (which is around the corner where NSW, Queensland and South Australia meet), but not in other anastomosing rivers with the ability for active scour. This is attributed to the alluvial history of the area, in particular the presence of a more easily-eroded sand sheet at relatively shallow depths beneath cohesive surface sediments (Silcock, 2009). The formation of waterholes can be split into three broad phases: initiation, augmentation and maintenance (Knighton and Nanson, 1994b).

During flow events, the waterholes retain water and act as a storage facility. They undergo continuous water loss by evaporation (Hamilton et al., 2005). Open water evaporation rates can be variable between waterholes, in part due to their different sizes, the effective fetch for wind action, the height and width of riparian vegetation and the degree of channel incision below the levees, which affects wind-induced turbulence at the water surface, exposure of the water surface to solar heating and convective air circulation above the water surface (Brutsaert, 1982).

Although waterholes are considered to be losing systems, available information in the literature presents contrasting hypotheses regarding the importance of groundwater inputs to waterholes in dryland regions like Cooper Creek (Qld). In this region, the waterholes range in persistence from those that dry up every few years to those, according to local and landholder knowledge, that have not been dry since Europeans settled in the area in the late 1800s (Davis, 2002).

Waterholes are also common features in other arid and semi-arid regions of the world. McLachlan and Cantrell (1980) defined three types of waterholes in Africa. They differ mainly in depth, which causes a gradation in life span following rain. The relative life spans of these waterholes extend from a few hours up to six months. In contrast, some of the waterholes mapped in Queensland appear to have permanent feature, for example the Georgina River is usually dry, with 16 permanent water holes along its course, which includes the Paravituri waterhole, located towards the eastern side of the Landsat image shown in Figure 2-3. During flood events the Georgina River transforms into anastomosing channels with width more than 25 kilometres.

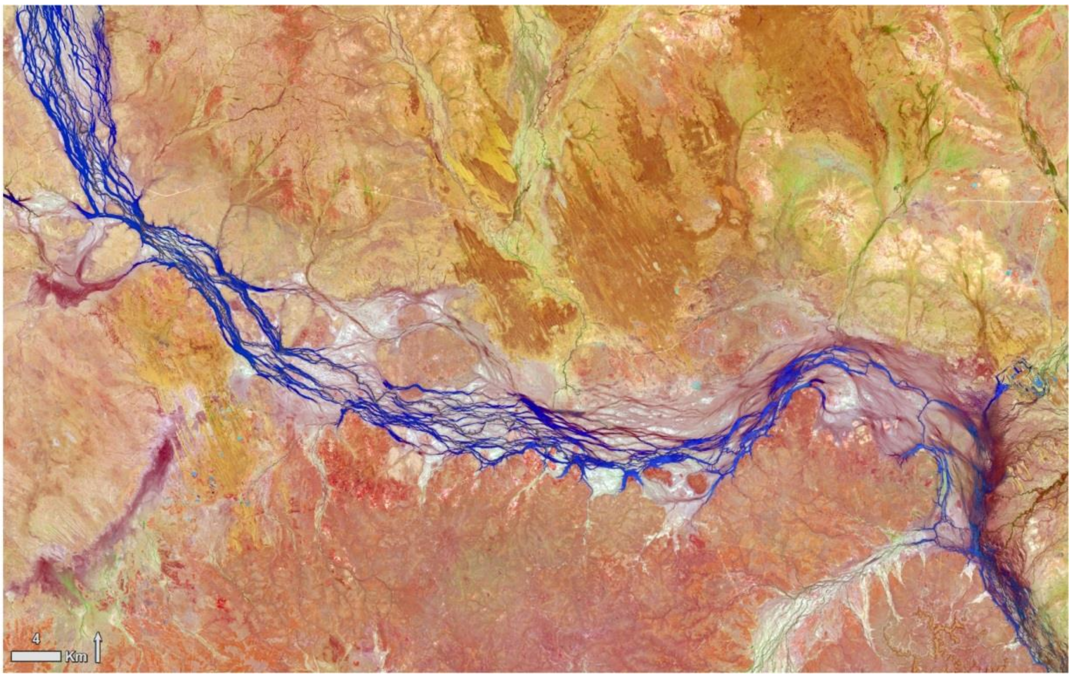


Figure 2-3. Anastomosing channel systems in the Georgina River, 19 January 2015 (After, Remote Sensing Centre, QLD).

Recently, in order to assist pastoral communities in the Horn of Africa, a consortium of agencies (NASA, USGS, universities, FAO, etc.) has established a real time monitoring system of waterholes in the Horn of Africa. Easy access to waterhole data can be obtained via the internet [3]. Satellite-based rainfall, modelled run-off and evapotranspiration data were used to model daily water level fluctuations. Mapping of waterholes was achieved with 97% accuracy (Senay et al., 2013). Validation of modelled water levels with field-installed gauge data demonstrated the ability of the model to capture the seasonal patterns and variations (Senay et al., 2013).

2.3.1 Evaporation and Evapotranspiration

Evapotranspiration is the second largest component of the water balance equation. Direct evaporation may take place from the stream channel or from billabongs and man-made storages. Annual potential evaporation values close to 4000 mm have been recorded in the central areas of Australia, which have exceptionally long hours of sunlight (Figure 2-4).

Evapotranspiration is the key component linking water balance and energy balance (Table 2-1). The primary factors controlling the long-term mean evapotranspiration are the local interaction of water supply (precipitation) and demand (potential evapotranspiration). Budyko (1974) assumed that actual evapotranspiration is controlled by both water and energy availabilities and considered that the ratio of annual evapotranspiration and annual precipitation is the function of the ratio of annual precipitation and annual net radiation. Based on this method, catchments or regions are divided into "water limited" and "energy limited" systems (Budyko, 1974).

[3] http://watermon.tamu.edu/

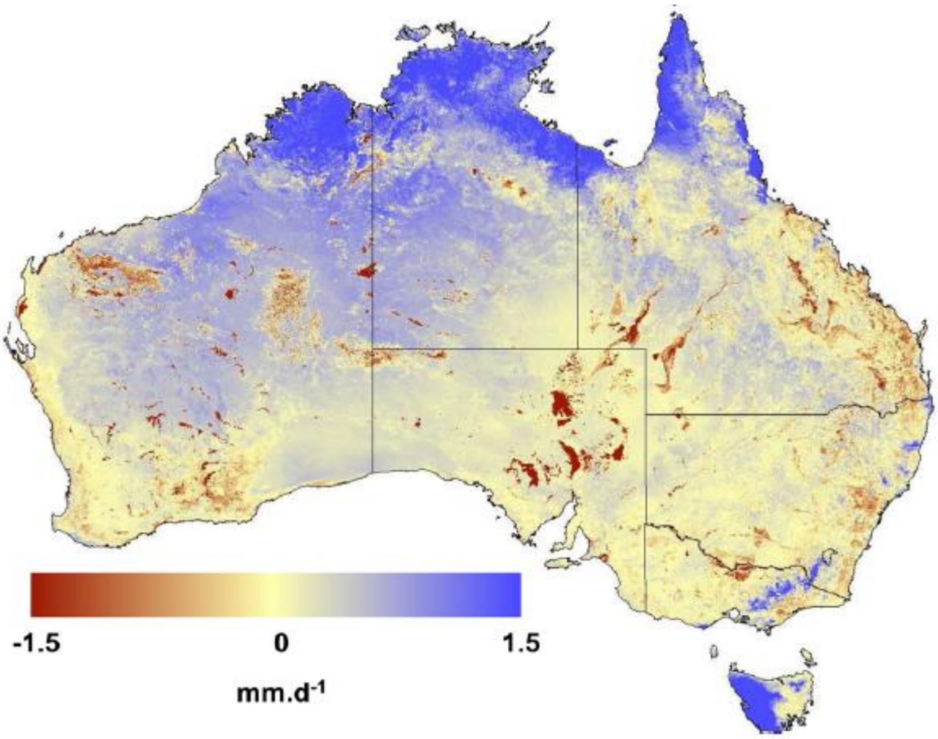


Figure 2-4. Mean P - ET water balance map for the entire Australian continent. The value was positive in most of the north of the continent, in western Tasmania and in the mountain ranges of SE New South Wales and Victoria. It was highly negative in all lakes, wetlands and irrigated areas, which evaporate, in addition to water received as precipitation, water from surface runoff and diversions (Guerschman et al., 2009)

Table 2-1: Average annual rainfall, surface runoff, groundwater 'recharge', and evapotranspiration for drainage basins in Australia (DPIE, 1997).

Drainage	Rainfall (P) (mm)	Surface Runoff (mm)	Evapotrans-piration (ET) (mm)	Groundwater Recharge (mm)	ET/P (%)
North-east coast	905	186	713	6	79
South-east coast	909	153	744	12	82
Tasmania	1352	776	568	8	42
Murray-Darling	472	23	446	3	95
South Australian Gulf	353	11	337	5	96
South –west Coast	459	21	428	10	93
Indian Ocean	259	8	248	3	96
Timor Sea	871	148	712	11	82
Gulf of Carpentaria	805	144	656	5	82
Lake Eyre	238	5	231	2	97
Bulloo-Bancannia	204	11	191	2	94
Western Plateau	262	1	260	1	99

Satellite based methods would have provided snapshots of evapotranspiration at continental and basin scales. Unfortunately, these data sets were not available for inclusion in this thesis.

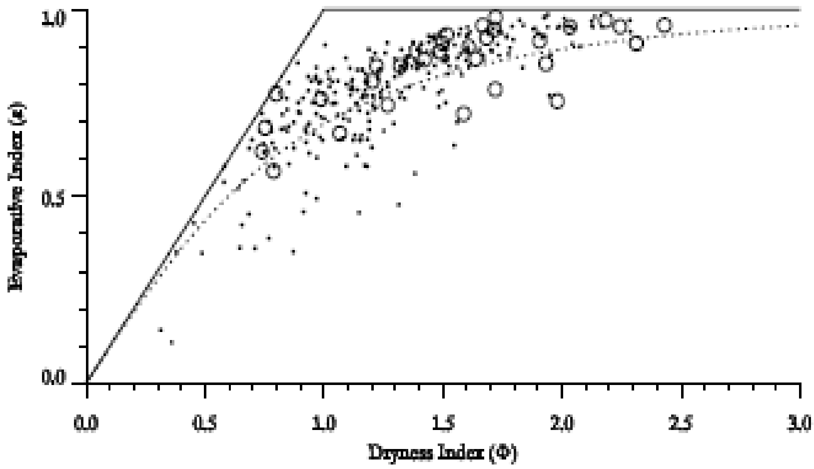


Figure 2-5. Plot of mass balance data from 331 Australian catchments showing the deviations of values around the Budyko curve. Large, hollow circles denote the 30 moderate-sized catchments (Ac>1000 km2) and small circles denote the remaining 301 smaller catchments (<1000 km2). Data are from Peel et al. (2000) and Raupach et al. (2001)

In the last decade, partitioning of total evaporation into soil evaporation and transpiration was carried out using stable isotopes (Rothfuss et al., 2010; Williams et al., 2004; Yepez et al., 2005; Zhang et al., 2010). Thus, the stable isotopic composition (18O and 2H) of ecosystem water pools such as leaf, root and surface water and atmospheric water vapour may help resolve some of the water requirements for GDEs. Fractionation processes lead to distinct isotopic signatures of water pools (Lai et al., 2006; Thorburn, 1994). Variations in the isotopic composition of water in different compartments (groundwater, soil, and vegetation) can provide information about water sources used by GDEs.

In order to determine persistence of aquatic refugia between flow pulses in the Cooper Creek catchment of western Queensland, isotopic and solute chemical tracers were used (Hamilton et al., 2005). This study concluded that the hydrology of waterholes in the Cooper Creek system is dominated by evaporative water loss (Hamilton et al., 2005). Given the complexity of hydrology and geomorphology of arid and semi-arid regions of this part of the world, the findings from this study should not be extrapolated to other waterholes in the region without site-specific knowledge on the hydrology (Hamilton et al., 2005).

In contrast, Larsen (2012) inferred that the role of evaporation in the Cooper Creek catchment could be negligible based on chemical-hydrograph separation technique. The bulk of TL was considered to contribute to groundwater recharge via scoured streambeds. Given that the sampling strategy employed in this study was biased towards a quick flow component of a hydrograph, which underestimates the role of ET in one of the driest parts of Australia, the major conclusions on the magnitude of recharge can be questioned.

2.3.2 Infiltration

Infiltration is the seepage of water through the stream channel, banks and flood plain. Infiltrating water can be viewed as potential recharge; it may become recharge, but it may instead be returned to the atmosphere by evapotranspiration (Healy, 2010). The rate of seepage depends on the type of channel material and the associated hydraulic properties of the channel material, channel geometry, wetted perimeter, circulating flow and sediments, and depth of groundwater. The magnitude and direction of flux depends on the following physical factors (Lerner et al., 1990b):

- Distribution of difference between stream and groundwater heads;
- Distribution of permeability of streambed and thickness;
- Permeability of aquifer;
- Geometric characteristics of aquifer;

- Geometric characteristics of the streambed.

Although a number of investigations have concentrated on infiltration processes in thick unsaturated zones linked to the transport of hazardous or radioactive wastes (Long and Ewing, 2004), only few have addressed water resources management issues related to focused groundwater recharge (Flint et al., 2000; Kulongoski and Izbicki, 2008; McCord et al., 1997). In a disconnected system, where the groundwater body is deep, most of the infiltrated water can be lost in the unsaturated zone on its way to the deep aquifer system (Phillips and Walvoord[4]).

One interesting case study is from the Mojave Desert in the US, where multiple methods (hydraulic, thermal, chemical and isotopic) were used to study recharge processes in a deep aquifer systems (Izibick, 2008). The unsaturated zone can be as much as several hundred metres thick in some places. It may also be composed of many different layers having different hydraulic properties that ultimately control the rate of downward movement and lateral spreading, and the flow of water through the unsaturated zone. In some reaches, streambed infiltrated water didn't reach the groundwater body. On the basis of this study, infiltration of about 0.7 m/yr represents a threshold, below which deep infiltration and subsequent groundwater recharge do not occur (Izbicki et al., 2008).

Most of the infiltration case studies in Australia were based on off-channel sites. Off-channel infiltration rates may differ significantly from in-channel infiltration rates. However, recently an attempt has been made to quantify focused recharge in the Namoi Valley (Berhane, 2006;

[4] http://web.sahra.arizona.edu/research[4], accessed on 13/10/2012

Brownbil et al., 2011) and in the Lake Eyre Basin (Costelloe et al., 2003; McMahon, 2008). Streambed infiltration rates reported in literature vary over a wide range. For the Negev desert in Israel, infiltration rates were estimated to be between 40 and 400 mm/h (Lange et al., 1999); for experimental research site in Abu-Dahbi, TL ranged 46–285 mm/h for a model input Osterkamp et al. (1995). In the Southwestern US, infiltration rates were calculated using an array of temperature sensors (Hoffmann et al., 2007). Reported infiltration rates were as large as 166 m/d, during the onset of streamflow. However, in this particular case study saturation of the unsaturated sediments reached in less than 10 minutes (Hoffmann et al., 2007).

Some studies have focussed on understanding the hydrological processes of discharge/recharge in arid regions of Australia. For example, the Cooper Creek system is mapped as a "groundwater-dependent ecosystem" (Hatton and Evans, 1998). This would seem to be based on the observation by local residents suggesting water levels in the waterholes often do not synchronise with the annual evaporation rate observed in open pans. On the other hand, the floodplain sedimentology, reviewed by Gibling et al. (1988) suggested that groundwater exchange is not significant. The waterholes lie on several metres of very fine clay that would tend to seal their basins against groundwater exchange, and an excavation to study the floodplain sedimentology in the southern Cooper Creek region showed that sand deposits beneath the clay bottom of a waterhole were dry. Costelloe et al. (2007) monitored surface water levels in a variety of waterholes across the Lake Eyre basin and concluded that there is no evidence of sustained connection between waterholes and regional shallow aquifers. It is also unlikely that the generally highly saline (TDS > 38,000 mg/L) regional shallow groundwater would be discharged on a large scale to waterholes given the depth of the groundwater table is more than 4–5 m below the base of most waterholes (Costelloe et al., 2007). Furthermore, there is a well-documented dependence of these ecosystems on fresh water (Bunn et al., 2006; Kingsford et al., 1999).

Based on hydrological investigations carried out in the Cooper Creek region, the flood-induced bed scouring process was suggested to enhance connectivity between waterholes and local groundwater systems (Cendón et al., 2010; Knighton and Nanson, 1994b), due to the presence of a thick clay layer. However, distributed recharge at a regional scale is not feasible (Cendón et al., 2010). Similar impedance would be expected around Cooper Creek where there are extensive floodplain and channel muds (Fagan and Nanson, 2004; Maroulis et al., 2007). In addition, in the Chookoo dune-floodplain complex, sand dunes stratigraphically connected to the underlying fluvial sand aquifers may provide another mechanism of groundwater recharge (Cendón et al., 2010).

In the northern MDB, large areas are underlain by heavy clay soils. In these catchments quantification of deep drainage, related to dryland salinity projects, was the main focus of numerous investigations (Cartwright et al., 2007; Silburn et al., 2011). Given the hydro(geo)logical complexity, it is a challenge to estimate recharge to deep aquifer systems overlain by a thick unsaturated zone, consisting predominantly of aquitard layers. For simplicity, in most of the dryland salinity studies, deep drainage was assumed to equal recharge. However, in a recent study carried out in the Lower Namoi Valley (Timms et al., 2011) an attempt was made to differentiate between deep drainage and recharge by comparing groundwater level responses to model predicted deep drainage over a 39 year period. During the study period groundwater levels were unresponsive to major rainfall events and most piezometers at about 18 m depth remained dry. This finding is in agreement with EIS reports carried out for coal mine development in the Namoi Valley[5].

2.3.3 Bank storage

Because of its complexity and difficulty of measurement, this component of the TL is the least studied in literature. During floods and high runoff events, water may infiltrate into the banks or flood plain. This water may be temporarily stored in the subsurface strata and seep back into the stream once the event has passed or during the recession of the event (Summerell et al., 2006). Riparian vegetation can rely on the water stored in bank storage and uses its deeper roots to tap water from groundwater during dry periods (SKM, 2011).

Gibson et al. (2008) applied stable isotopes of water data in combination with runoff discharge data, to make an initial evaluation of the ungauged gains and losses along the Darling River. They found that, by evaluating the physical monitoring data alone, the Darling River appeared to be a slow draining river with little exchange. Later, the inclusion of isotopic data in the evaluation revealed that the system is much more complex. It was surmised that several reaches of the river were volumetrically gaining, including the Bourke to Louth and Louth to Wilcannia reaches.

Hydrochemical and isotopic tracers, in conjunction with high resolution sampling strategies were applied, to quantify the contribution of evaporation, bank storage release and saline

[5] www.whitehavencoal.com.au/.../67417Part2_GroundwaterReport.pdf accessed in November 2012

groundwater influx to the evolution of the Darling river waters by (Meredith et al., 2009). Fractional contributions (% of volume) of groundwater to the river water were calculated for different reaches, using Cl concentrations, d18O and d2H values, and it was found that river waters comprised of approximately 60–99% saline groundwater during zero flow. This high salinity water was found to dilute back to lower values before reaching Louth because of the input of fresh, isotopically depleted waters that mix into this stretch of the Darling River. These waters are most likely to represent the release of bank storage during low-flow periods (Meredith et al., 2009).

It appears this study has not resolved the issue of surface groundwater connectivity on the targeted reaches of the Darling River. During drought the channel appears to be losing. During the wet cycle, however, bank storage appears to be contributing saline groundwater to the Darling. The paper doesn’t provide the magnitude of discharge in proportion to recharge or stream losses. The ambiguity may have been resolved by integrating hydrogeological data with hydrochemical information. In addition, the Darling River catchment is complex, none of the investigators assessed the role of perched systems and deep laying groundwater system.

2.4 Methods to determine TL

Transmission losses have been studied in different catchments and climatic settings in Australia, South Africa, USA, Israel, and Saudi Arabia (Abdulrazzak and Sorman, 1994; Cornish, 1961; Costelloe et al., 2003; Hughes, 1992; Jordan, 1977; K.E.Saxton, 1962; Knighton and Nanson, 1994a; Lane et al., 1980; McMahon, 2008; Sharp, 1962; Walters, 1990). Most of the studies were related to ephemeral streams. Commonly used methods of estimating TL have been reviewed by many authors (Cataldo et al., 2010; Costa et al., 2012;

Kinzelbach et al., 2002; Scanlon et al., 2006; Vivarelli and Perera, 2003). Some of the earlier reviews put more emphasis on engineering based methods. Techniques aimed at quantifying infiltration rate and subsequent recharge in ephemeral and intermittent streams rely on physical, chemical, and isotopic approaches (Kulongoski and Izbicki, 2008). A summary of methods reviewed is given in Table 2-2.

Table 2-2: Techniques for estimating transmission losses

Categories	Method	Scale	Advantages	Disadvantages	Reliability in terms of measurement errors
Statistical	• Regression equation	Reach	Simple to apply and works in infiltration dominated systems	A black box model;	Low
Physical	• Green-Ampt Approximation	Point		Works well in homogenous medium	Medium
	• Channel water budget	Reach	A physical based and simple to apply	Errors associated with measurements	High
	• Coupled surface/gw models	Catchment/ Sub-catchment	Simple to apply	Data hungry/Expensive	High-Medium
	• Hydrologic routing	Catchment	Ideal for scenarios development and management Prediction tools	Not user friendly	Medium
Heat	• Natural environmental tracers	Point/Catchment	Easy and affordable	Not an ideal tracer (Retardation)	Medium
		Point/Catchment	Expensive	Not an ideal tracer (Retardation)	High
Chemical	• Fiber optics	Point/Catchment	Complementary		Medium
	• Natural and radioactive			Expensive	
Isotopic		Point/Catchment			Medium
	• Stable isotopes	Point/Catchment			
	• Radioactive isotopes				

2.4.1 Simple Regression Equations

Based on a simple regression equation between inflow and outflow within a channel reach, relationships have been found between TL and inflow volume (Lane, 1971). Others explored non-linear relationships between targeted parameters and found that log transformed data of upstream flow volume and channel width have a significant relationship with TL (Sharma, 1992; Walters, 1990).

In Saudi Arabia, several regression equations were developed to estimate TL and recharge from wadi-beds (Sorman and Abdulrazzak, 1997). These studies indicated that the magnitude of TL and consequent recharge resulting from measured upstream runoff hydrographs is a function of stream and sediment characteristics (Cataldo et al., 2010).

In general, regression models as black box models are straightforward to use. However, the models are not linked to physical processes governing TL. For example, Walters (1990) provided three regression equations for estimating transmission losses in southwest Saudi Arabia. One of the equations was found to be a reliable predictor of losses associated with small upstream volumes. The other two equations can be utilised to make TL predictions for losses associated with large floods however the data set studied was relatively small (Walters, 1990), and a number of the variables rejected in this regression analysis (for example hydraulic conductivity, drainage area, and lateral inflow) may be important.

2.4.2 Hydrologic Routing

Knighton and Nanson (1994b), developed a three parameter Muskingum procedure to estimate outflow hydrographs and TL for the channel-country in the Lake Eyre Basin. The purpose of the study was to investigate the effects of local hydrologic variability on TL. The estimated parameters behaved well when the flow was within the primary channel. Predicted TL was less when clay sealing prevented TL over significant lengths of the river, and evaporation and drainage diffusion were major causes for TL. This is one of the reasons why a simple black box model fails to adjust to changing geomorphological conditions (Cataldo et al., 2010; Lange, 2005).

More recently, Morin et al. (2009) applied a simplification of the full one-dimensional St. Venant equations that combine continuity and momentum–conservation equations to estimate recharge in the Kuiseb River of Namibia, South West Africa (Figure 2-6). This study demonstrated when channels are short, narrow or have limited infiltration potential, flow duration controls infiltration rate. On the other side of the infiltration spectrum, for channels that are long or wide, or have high infiltration rates, the opportunity for infiltration is higher and the flood magnitude (i.e., peak discharge) plays a more important role in determining infiltration volumes (Morin et al., 2009).

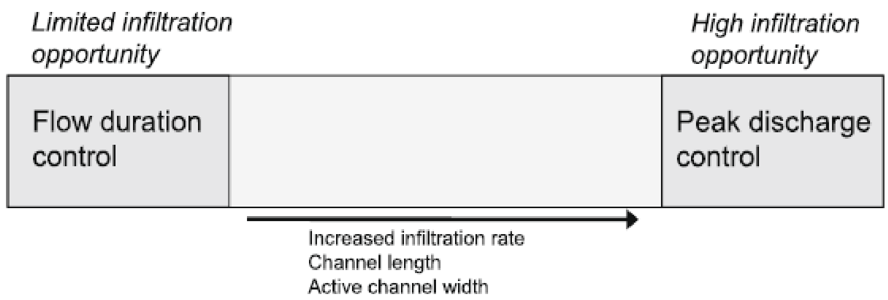


Figure 2-6. Scheme presenting the importance of flood duration and peak discharge on infiltration volume for different channel characteristics (after Morin et al., 2009)

2.4.3 The Green-Ampt Approximation

Reid and Dreiss (1990), demonstrated that, compared to numerical solutions, the Green-Ampt model (Figure 2-7) performs well in representing infiltration from ephemeral streams. The Green-Ampt equation was developed using Darcy's Law in conjunction with geometry and hydraulic properties (Eq. 2-1). The rate of infiltration is approximated by the following expression:

$$q = -K_s \frac{dh}{dz} = -K_s \frac{h_2 - h_1}{z_2 - z_1} = -K_s \frac{(\psi_f + z_f) - (H + 0)}{z_f - 0} = -K_s \frac{\psi_f + z_f - H}{z_f} \qquad \text{Eq. 2-1}$$

where:

H = the depth of ponding, cm;

K_S = saturated hydraulic conductivity (cm/s);

Q = flux at the surface (cm/h) and is negative,

Ψ_f = suction at wetting front (negative pressure head);

Θ_i = initial moisture content (dimensionless) and

θ_S = saturated moisture content (dimensionless).

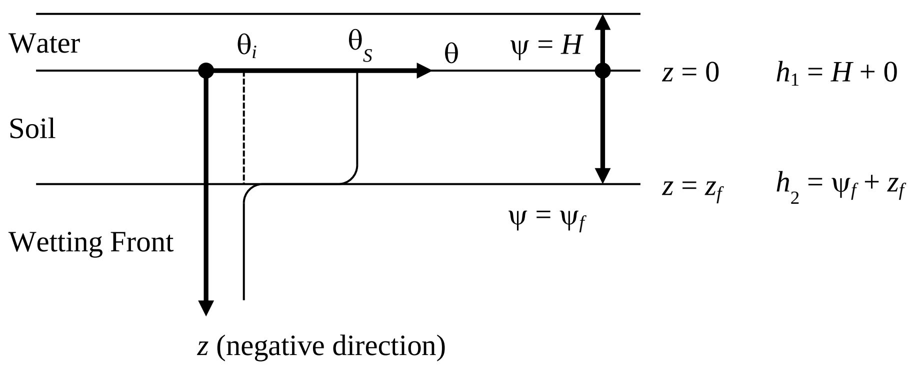


Figure 2-7. Schematic diagram of wetting front movement in the unsaturated zone, as conceptualised in the Green-Ampt model (after Hornberger et al., 1999)

2.4.4 Channel-Water Budget/ Coupled Surface Groundwater Models

A Channel water budget approach is based on the volume difference between two consecutive gauging stations along a stream channel and estimating the maximum infiltration rate based on a simple mass balance calculation (Lange, 2009; Shentsis et al., 1999).

The range of infiltration rates that can be measured using this technique depends on the magnitude of the TL relative to the uncertainties in the gauging data and tributary flows. Lerner et al. (1990b) reported that measurement errors during high flows are often ±25% and can range from –50% to +100% during flash floods in semi-arid regions. The recharge estimates represent average values over the reach between the gauging stations.

During the last two decades, coupled surface and groundwater models are used to study complex interactions between surface and groundwater (MIKE SHE, MODHMS MODFLOW, HydroGeoSphere, and GSFLOW). However, all models show some limitations when it comes mimicking disconnected river systems.

2.4.5 Heat Tracer

Vertical temperature profiles can be used to estimate infiltration rates according to Bredehoeft and Papadopulos (1965b). Generally, downward heat transport is controlled by conductive transport during a dry spell and by advective transport mechanisms during channel infiltration or exfiltration. In surface water dominated catchments, deviations of purely conductive propagation of temperature fluctuations from the land surface into the subsurface suggest the occurrence of infiltration (Rorabaugh, 1954; Stonestorm and Constantz, 2003). Usually, streambed temperature measured at different depths is analysed to estimate fluxes and hydraulic conductivities, using numerical techniques (e.g., VS2DHI, FEFLOW and SUTRA) or analytical methods (eg, EX-STREAM, VFLUX).

2.4.6 Chemical and Isotopic Methods

The chemical and isotopic signatures from rainwater, surface water and groundwater can be used to develop mixing ratios between different components of the water balance equation (Cendón et al., 2010). Deuterium compositions in downstream direction of the River Murray tend to increase due to evaporation (Simpson and Herczeg, 1991a).

In the context of the study of TL, the meteoric water line provides an important key to the interpretation of deuterium and oxygen-18 data (Hamilton et al., 2005). Deuterium (δ2H) and Oxygen-18 (δ18O) are the two commonly used stable isotopes to study hydrological processes (Evaporative loss and mixing). When Deuterium and Oxygen-18 signatures of rainwater across the world are plotted they produce a reference-straight line, called the Global Meteoric Water Line (GMWL), the equation for the GMWL is $\delta 2H = 8\delta 18O + 10‰$ (Craig, 1961). For example, the meteoric water line of rainfall samples from the Amazon (humid climate) is represented by an equation similar to the GMWL: $\delta 2H = 8\delta 18O + 11‰$. While rainfall samples from an arid region in the Darling catchment is represented by an equation of the form: $\delta 2H = 5\delta 18O - 2‰$ (Craig, 1961).

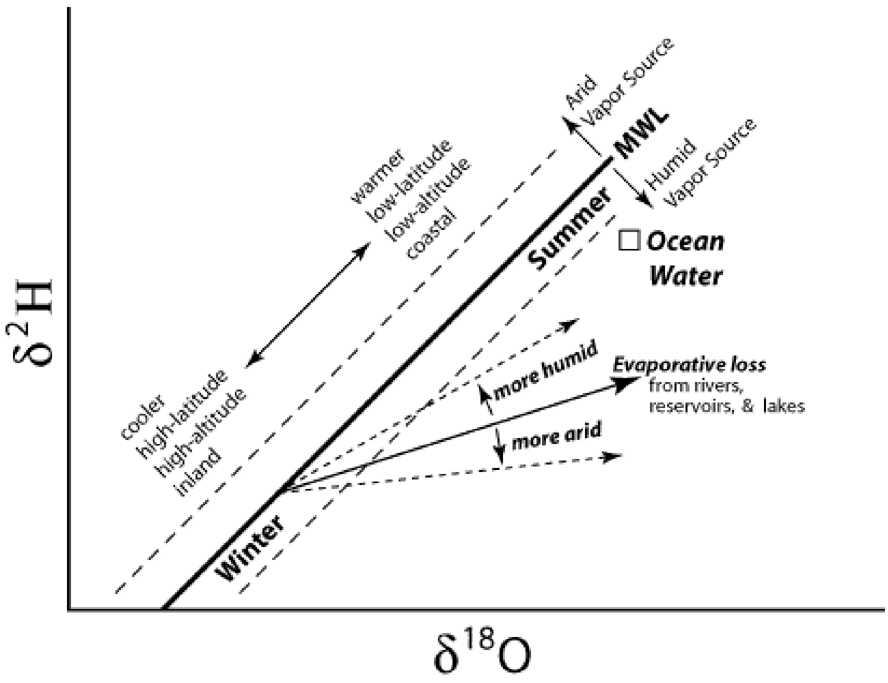


Figure 2-8. Deviations in isotopic compositions away from the meteoric water line as determined by climatic conditions (SAHRA webpage accessed on 12/06/2012).

Water with an isotopic composition falling on the meteoric water line is assumed to have originated from the atmosphere and to be an unaffected by other isotopic processes (Domenico and Schwartz, 1998).

Due to river induced recharge, the isotopic composition of groundwater may change based on end member isotopic compositions of surface and groundwaters. Depending on the availability stable isotope time series data, it is possible to develop a simple mixing model between surface and ground waters using a mass-balance approach (Christophersen and Hooper, 2012; Clark and Fritz, 1997; Kendall and McDonnell, 1998).

Chemical and isotopic methods have been successfully applied for estimation TL losses in the MDB (Simpson and Herczeg, 1991b). For example, having a good understanding of the geological environment, Radon can be useful tracer for estimation infiltration rate.

2.4.7 Use of remote sensing for estimation ET

The assessment of channel TL is traditionally carried out by doing a streamflow water balance and comparison between streamflow and groundwater levels (Lerner et al., 1990a). Satellite data might contain complementary information as they have been proven to be a valuable resource to improve understanding of surface hydrological processes (Van Dijk., 2011). For instance, remotely sensed thermal infrared imagery collected by Landsat provides estimates of land-surface temperature that allow mapping of evapotranspiration (ET) at the spatial scales at which water is being used (Anderson et al., 2012).

In a recent study, to assess groundwater dependent ecosystems in the Namoi Valley (SKM, 2011), the simulated ETa rate decreased from upland to low land areas using the Surface Energy Balance Algorithm for Land (SEBAL) an algorithm developed by WaterWatch. The primary basis of the SEBAL model, which was developed and reported by (Bastiaanssen et al., 1998), is the surface energy balance. Within the model evapotranspiration is calculated as the residual of the surface energy balance equation:

$$ET = Rn - G - H \qquad \text{Eq. 2-2}$$

where:

ET = latent heat flux (W/m2),

Rn = net radiation flux at the surface (W/m2),

G = soil heat flux (W/m2) and

H = sensible heat flux to the air (W/m2) as shown in Figure 2-9.

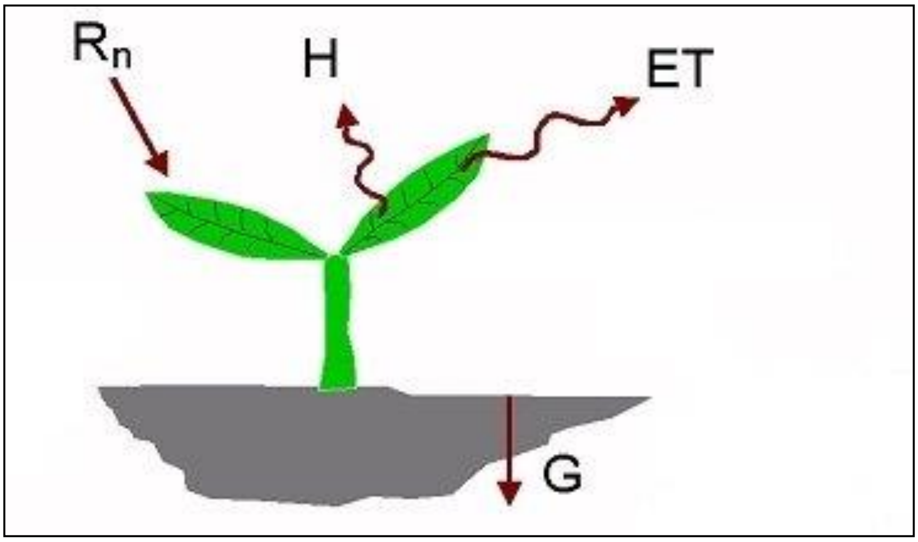


Figure 2-9. Components of the energy balance (SEBAL)

The methodology provides an insight into the magnitude of ETa at a catchment scale. In the near future, satellite derived consumptive use information at a catchment scale is expected to be routinely applied for water resources assessment at different spatial and time scales depending on the types of satellite used (SKM, 2011).

Although of low resolution, ET can also be derived using MODIS satellite, which was first launched aboard the Terra satellite in 2000. The full 12-year evapotranspiration time series is available as 8-day, monthly and annual averages. A toolbox for downloading monthly MODIS data is available in the ArcGIS Resource Center. Recently, researchers from the University of Montana have made available the average annual evapotranspiration for the period of record (2000-2015). Availability of ET products in conjunction with other hydrological data is expected to facilitate estimation of ET and TL at a catchment scale.

2.4.8 Gravity Survey

Time series microgravity surveys have been used to monitor changes in subsurface water storage and recharge near ephemeral channels in Arizona (Pool and Eychane, 1995; Pool and Schmidt, 1997). The method measures changes in subsurface mass by applying Newton's law of gravitation to variations in the results of repeated gravity surveys from a network of stations.

Space-borne techniques can provide information on water storage changes on a regional and global scale. For instance, (Leblanc et al., 2009) performed a study of the multi-year drought in the MDB quantifying the propagation of water deficits through the hydrological cycle. They found a high correlation between the observed groundwater variations from boreholes and the Gravity Recovery and Climate Experiment (GRACE) satellites signals when the long recent drought in Australia (2000–2009) had reduced the available surface water resources. The net loss of water over the period of GRACE observations (2002–2007) was found to be about 200 km3.

In similar regional water resources assessment study carried out in the Middle East, researchers at the University of California, Irvine, NASA and the National Centre for Atmospheric Research have identified the Tigris and Euphrates River Basin having the second fastest rate of groundwater storage loss after India (Voss et al., 2013).

GRACE data show an alarming rate of decrease in total water storage of approximately -27.2 ± 0.6 mm yr^{-1} equivalent water height, equal to a volume of 143.6 km3 during the course of the study period. Additional remote-sensing information and output from land surface models were used to identify that groundwater losses are the major source of this trend (Voss et al., 2013).

2.5 Case Studies of TL in Australia

McMahon (1992) collected a worldwide database of annual runoff data for over 900 catchments. These vary in size from less than 10 km2 to in excess of 106 km2 and represent all the major world climate regions (Smith, 1998). Based on this study, Australia is not only the driest inhabited continent, but also has the largest inter-annual variability in rainfall.

In a first of its kind, Smith (1998) published water balance components for Australia at drainage scale. The catchments display contrasting characteristics. The catchment of the Mulgrave River, close to Cairns, is the wettest of all, with average annual rainfall of 3196 mm, runoff of 2074 mm and evaporation loss of 35%. At the other extreme, the driest catchment is Lake Frome in the Lake Eyre drainage division, which has an average rainfall of 156 mm of which 98 mm is lost to evapotranspiration (Smith, 1998).

The water balance components, presented earlier in Table 2-1, provide some indication on the magnitude of TL, where a big proportion consists of evapotranspiration. In this review, case studies on TL are presented from two areas in Australia: Part of the Lake Eyre Basin in Queensland, representing an arid region of Australia and the Murray Basin in NSW, representing a semi-arid region. These two regions are characterised by high ET/P ratio, greater than 90% (Table 2-1).

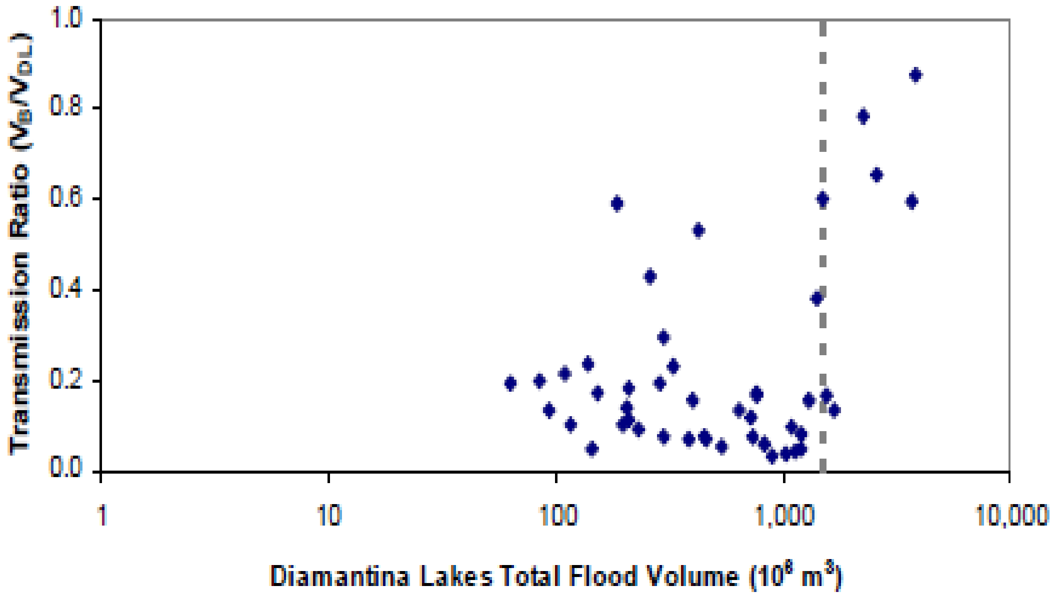


Figure 2-10. Transmission losses for Diamantina Lakes to Birdsville reach (Costelloe et al., 2006)

2.5.1 Lake Eyre Basin

The Lake Eyre Bain (LEB) covers one sixth of the Australian continent. On the global scale, the LEB is one of the largest internally draining systems; has the fifth largest terminal lake; is in the driest inhabited continent; and is drained by the most variable, major river systems, namely the Georgina, Diamantina and Cooper[6] (Puckridge et al., 1988). Potential evaporation exceeds actual evaporation in all parts of the LEB, except in the very headwaters of Cooper Creek (Ceplecha, 1971).

[6] http://www.ilec.or.jp/database/database_old.html: , accessed on 25/11/2012

In the upper reaches of LEB river systems, a linear relationship exists between annual discharge and catchment area, but in the middle to lower reaches the mean annual discharge decreases with increasing catchment area (McMahon et al., 2005). The transition from a positive relationship between flood discharge and catchment area to a negative relationship in the major rivers of the LEB coincides with the junctions of the major tributaries of the rivers. Downstream of these junctions, the river/floodplain width widens dramatically (known as the ‘Channel Country’(McMahon et al., 2005) and their floodplains are usually between 10 and 70 km wide (DERM internal report). They flow through semi-arid to arid climate zones and are ephemeral rivers, which means that the rivers are dry for most of the year and have water flowing through them only after there has been enough rainfall in their catchments. This rainfall usually occurs during summer and is related to monsoonal weather systems and sometimes cyclones in tropical northern Queensland.

In comparison to other basins in Australia, TL was studied in ‘detail’ on the lower reaches of three river systems of the LEB. The TL behaviour of flood events for the Diamantina Lakes to Birdsville reach of the Diamantina River is shown in Figure 2-10. The TL varied non-linearly with flood size and generally declined as the flood event volume increases towards a threshold value (approximately 1500 x 106 m3 for the Diamantina reach where flood pulses experience up to 96% TL) before increasing at flow volumes above the threshold value of the Diamantina River (Costelloe et al., 2003; McMahon et al., 2005).

Given the importance of TL in arid and semi-arid regions of Australia, the next logical hydrological issue would be assessment of recharge or what proportion of TL is transformed to groundwater recharge in the LEB. Considerable research has examined transmission losses and recharge rates of ephemeral streams with predominantly coarse grained channel sediments. However, far less is known of the recharge rates from rivers with corresponding fine grained channel and floodplain sediments that are more typical of many Australian arid zone rivers (Costelloe et al., 2012b).

Multiple methods were used to assess localised groundwater recharge during flood events in an ephemeral river in the Lake Eyre Basin. The study focused on the flood plains aquifers near two semi-permanent water bodies, Ultoomurra and Yelpawaralinna Waterholes in the lower reaches of the Diamantina River, which is known locally as Warburton Creek (Irvine et al., 2006). The results from this investigation showed changes in groundwater responses without any changes in groundwater quality or isotopic signatures, suggesting that for the period of monitoring the magnitude of recharge is negligible.

Quantification or assessment of recharge and discharge processes in the Lake Eyre Basin is complex due to the fact that upward fluxes from the GAB aquifer systems and downward fluxes induced by stream leakage as a component of TL can take place in waterholes and wetlands. A recent case study from the LEB (Costelloe et al., 2012b) shows how stable isotopes in conjunction with hydrochemistry could be used to distinguish between recent pluvial/fluvial recharge from an ancient artesian groundwater discharge. Mixing models, combined with forward evapo-concentration models, confirmed that the contribution of artesian groundwater was minimal in areas of Na-Cl-SO4 unconfined groundwater underlain by Na-HCO3-Cl artesian groundwater (Costelloe et al., 2012b). In contrast, based on stable isotope data, the recharge mechanisms for the GAB artesian groundwater(s) and shallow unconfined groundwater were considered to be similar (Tweed et al., 2011). Probably, the isotopic signal considered in this case study is a mixture of numerous wet and dry climatic periods making it difficult to resolve isotopic signatures for use in water resources management (Herczeg and Leaney, 2010).

2.5.2 Murray Darling Basin

The MDB covers about 15% of south eastern Australia, approximately 1 million square kilometres. Two million people live in the Basin and are dependent on it for their drinking water, including 1.2 million residents of the city of Adelaide (Craik and Cleaver, 2008b). Long term average rainfall in the Basin is approximately 500,000 GL per annum, yet the vast majority of this evaporates. Average annual runoff is 24,300 GL with 11,400 GL of long term average diversions (Craik and Cleaver, 2008b). As shown in Table 2-3, ET constitutes 95% of the water balance equation of the MDB.

There are significant differences in rainfall and inflow reliability between the northern (Darling River) and southern (Murray River) parts of the Basin. These differences have influenced irrigation development, water use and planning. Historically, the most reliable rainfall in the Basin occurs in the alpine regions of the Southeast, which supplies the Murray River system (Craik and Cleaver, 2008a). The Murray River is highly regulated, with over $2b USD of infrastructure. The two largest dams are Hume Dam (3033GL) and Dartmouth Dam (3905GL). The historically high reliability of inflows into these dams has supported the creation of very high reliability irrigation water entitlements and supplied urban and domestic water supplies throughout the Southern Basin (Craik and Cleaver, 2008a).

Table 2-3 Water balance for the Murray-Darling Bain (Peel et al., 2004 and UNESCO, 2004)

	Depth (mm)	Volume (GL)	%
Rain	480	508000	100
Evapotranspiration	457	-484000	95
Transfers from other catchments (from the Snowy and Glenelg rivers)	0.9	1000	0.2
Runoff	24	25000	5
Runoff is consumed as			
Evaporation from open water	2.8	-3000	0.6
Evaporation from flood plains and wetlands	7.6	-8000	1.6
Irrigation diversions	10.4	-11000	2.2
Discharge at the mouth of the River Murray	2.8	-3000	0.6

The new MDB plan came legally into effect on 24 November 2012. The Basin Plan includes enforceable limits on the quantities of surface water and groundwater that can be taken from the MDB water resources. The sustainable diversion limits (SDLs) are set initially at 2,750 gigalitres less than current diversions in the rivers across the Basin to be achieved by 2019 to provide additional water for the environment.

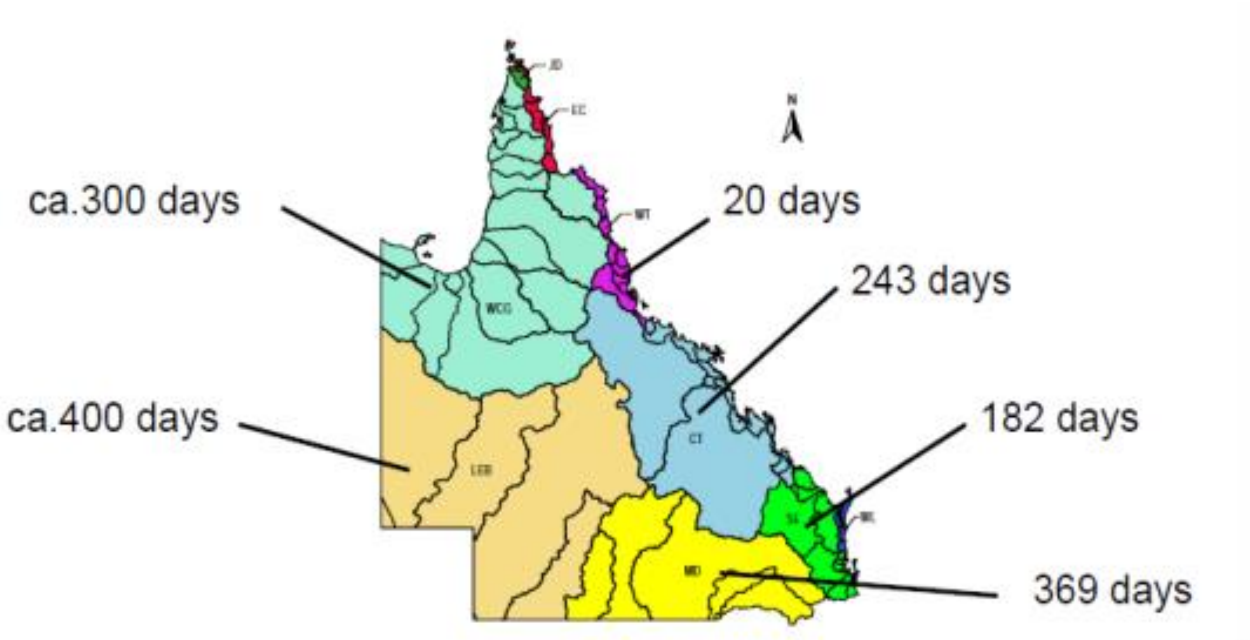


Figure 2-11. Regional average maximum no-flow spell duration (Marshall, 2006)

2.5.3 Northern MDB

The northern MDB includes seven catchments (Condamine Balonne, Paroo Warrego, Namoi, Border Rivers, Macquarie Castlereagh, Barwon Darling, Gwydir), which have different hydrogeomorphological characteristics, when compared to their southern counterparts. The northern Basin is drier than the south, and also much flatter than the southern basin. The flatter landscape means the Barwon-Darling River has a very low gradient (or slope), leading to the formation of major wetland areas in the lower areas (MDB, 2012).

In the northern part of the MDB, no flow duration can extend for more than 370 days, which has a detrimental effect on the ecosystems (Marshall, 2006b). Even in the wettest part of Queensland, outside the MDB, rivers dry-up for an extended period ranging from 20 to 180 days (Marshall, 2006b). In this type of environment, knowledge of TL is a prerequisite for effective water resources management.

In terms of persistence of flow, the Namoi River shows similar statistics as the river systems in Queensland. For example, at Gunnedah gauging station (419001), the longest dry period of 252 days with zero flow occurred from 19th January to 27th September, 1902 (WRIC, 1970). In more recent times, the river ceased to flow for 18-month period between July 1994 and December 1995.

For Queensland catchments, Weeks (1978) developed an empirical equation to estimate TL. Later, Stewart (1984) made an attempt to estimate TL related to releases from Leslie Dam using different empirical equations. The calculated TL ranged from 5 to 55 % of the inflow volumes.

In addition, groundwater recharge, a component of TL, has been studied in the Condamine Catchment (Huxley, 1982; KCB, 2011; Lane, 1979; SKM, 1999). The last three reports were based on the most popular groundwater model, MODFLOW. This package however, may not necessarily mimic complex hydrological processes in disconnected groundwater systems Simpson and Herczeg (1991a). Rainfall recharge is estimated to vary from 0.1 to 1% of the gross precipitation (Barnett and Muller., 2008). If recharge exists, it is probably induced by stream leakage or takes place via upland fractured rocks.

2.5.4 Southern MDB

To estimate TL, Barma and Varley (2012) used the stable isotopes of water (18O and 2H) together with major ion chemistry to estimate salt loads, evaporation and the hydrochemical evolution of surface waters in the Murray River, which is part of the larger Murray-Darling Basin (MDB). For the Murray River system, they estimated that the cumulative evaporation losses along the length of the river were approximately 35% of the initial volume of surface water in the system. But, it is not clear from this study whether infiltration loss was accounted for in the mass balance calculation.

Current river system models typically lump all components of transmission loss into a single value, which is varied on an annual basis (Barma and Varley, 2012). This approach does not simulate the processes involved in TL and is not able to reproduce the variations that occur on a seasonal or daily basis. Given that TL can be of the same magnitude as low flows, this approach makes it difficult for current river system models to accurately represent low flows Thoms et al. (1999).

A hydrological study carried out for the southern reaches of the MDB reports losses of about 4000 GL/yr on average (Gippel (2006). It was also reported that some losses vary considerably between years, which is shown in Figure 2-12 (Gippel, 2006).

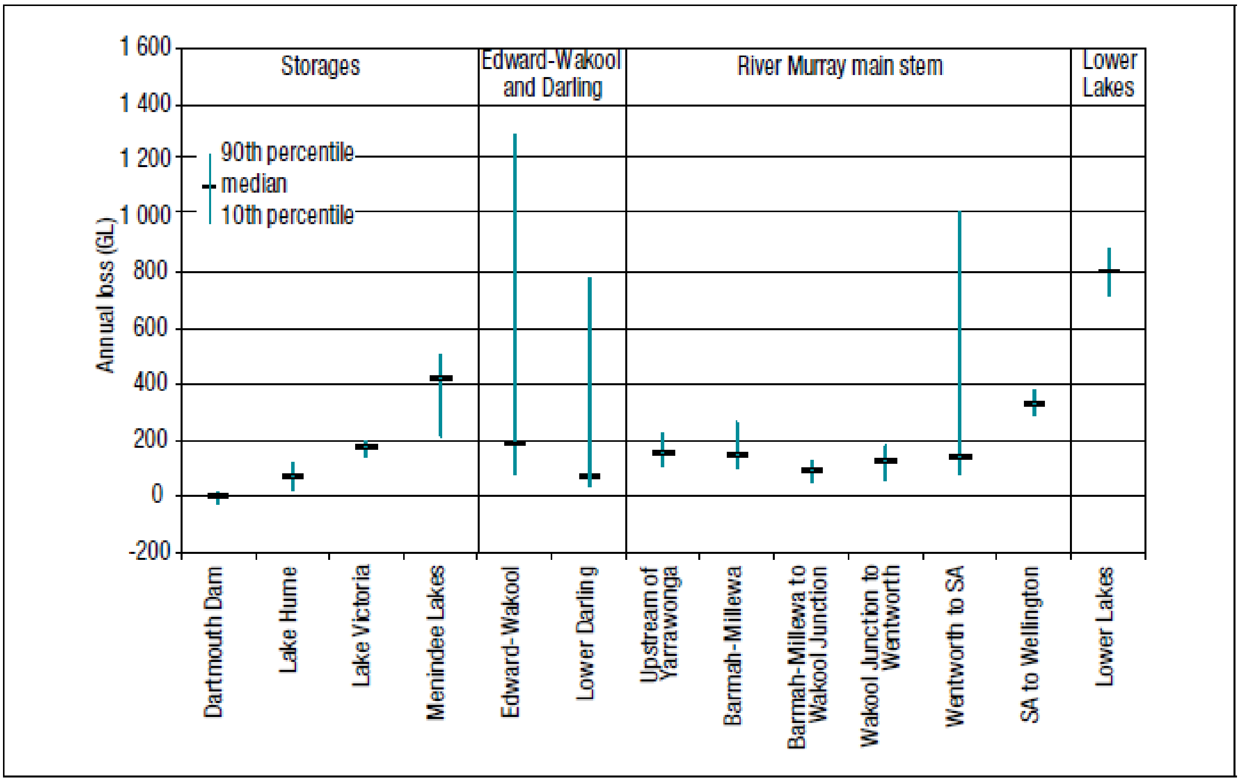


Figure 2-12. Variability in annual transmission losses in various river reaches in the Southern MDB (Source: (Gippel, 2006)

2.5.5 Namoi Valley

As the Namoi Valley is the catchment with highest groundwater development level within the MDB, it is appropriate to provide some background information regarding surface groundwater connectivity issues at different spatial scale.

At the catchment scale, Thoms et al. (1999) reported that the long-term average annual runoff for the Namoi River is 770,000 ML at Gunnedah, which represents 6% of the average annual catchment rainfall. Most of the runoff is generated in the headwaters of the valley (Cockburn, Peel and MacDonald catchments) about 90% of the total runoff comes from 40% of the catchment area. In the upland catchments annual flows increase with increasing catchment area, but downstream of Gunnedah annual flows decrease due to increasing evaporation, TL and water use for irrigation (Williams et al., 1989). The Lower Namoi was identified as a losing stream (Pigram, 1971).

A preliminary investigation to study the interactions between surface and groundwater in the Namoi Valley was carried out by Pigram (1971). He concluded that the three major streams in the Namoi Valley are basically losing water through bed infiltration. This is more likely with the Mooki River and Cox's creek, where flows are intermittent and ephemeral in nature. However, localised groundwater mounds are known to develop beneath the beds of these streams, following flash floods, dissipating slowly during ensuring dry periods (Pigram, 1971). Along the Namoi River some valley constriction near Gunnedah, coincident with the confluence of the Mooki and associated underflow of groundwater from the Breeza Plain, appears to convert a stretch of the river to effluent seepage (Gregory and Walling, 1973).

In the context of this review, the Namoi Valley has been divided into five sub-catchments. As a catchment is a basic unit in geomorphology (Ivkovic, 2009b; Ward, 1984; Winter et al., 1998), this subdivision is considered to be appropriate for water balance assessment. In turn for water resources management purposes, the Upper Namoi Valley was divided into 12 zones (Figure 2-13).

For reaches with poor hydrological data, the flow duration curve (FDC) has been used to characterise streams into gaining or losing (Ivkovic, 2009b; Winter et al., 1998) as a surrogate indicator. For example, for a flatter curve, the stream is considered to be supported by delayed runoff (interflow or 'baseflow'). A steeper curve is less likely to be supported by groundwater discharge. It should be noted that individual reaches might switch from gaining to losing conditions depending on the scale of investigation.

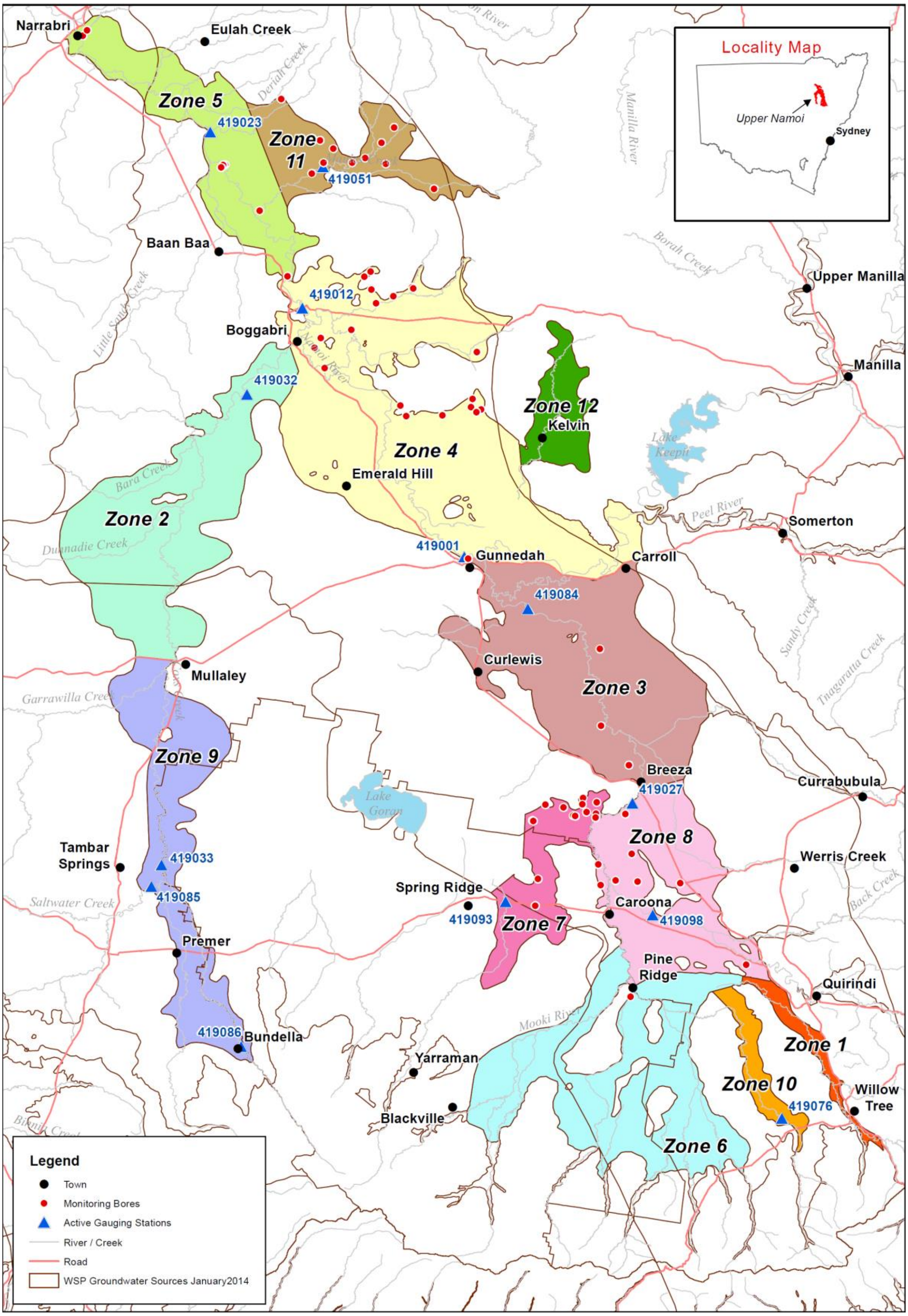


Figure 2-13. Upper Namoi groundwater source location map (Source: NOW, 2015).

2.5.6 Peel Sub-Catchment

The unregulated tributaries of Attunga Creek, Moore Creek and Duncans Creek are largely ephemeral: losing water to groundwater in times of high flow then gaining groundwater again until the level of the groundwater drops below the bed of the creeks resulting in the creeks drying up during dry times (O'Rourke, 2010). The Cockburn River is both a losing and gaining system depending on location. Goonoo Goonoo Creek is largely a groundwater-driven gaining stream in times of low flow. Dungowan Creek is typically a losing stream (Ward, 1984; WRC, 1970).

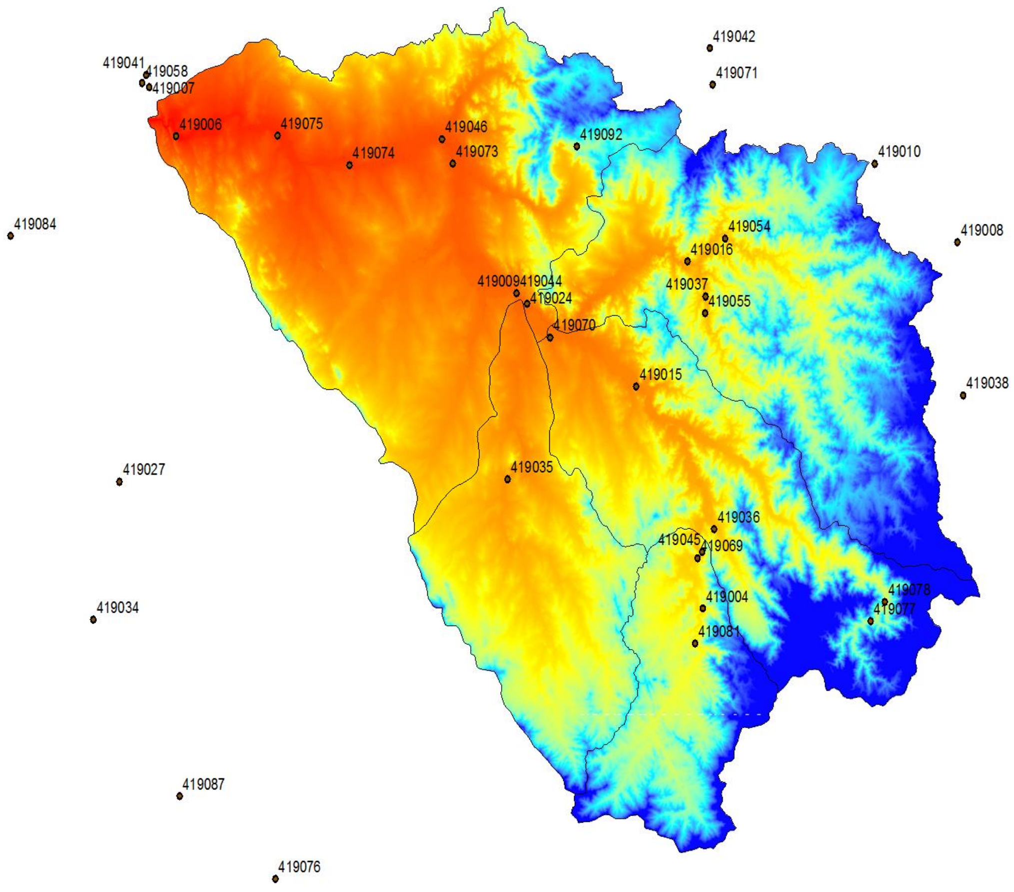


Figure 2-14. Regulated and unregulated sub-catchments of the Peel Valley (numbers shown in the map are gauging stations)

As part of the GDE study, (Berhane and Serov, 2004) delineated the reaches of the Peel River between the Chaffey Dam and the Carroll Gap. The Peel River upstream of the Tamworth gauging station was losing (Berhane, 2006). Due the absence of piezometers adjacent to gauging stations in the area between Tamworth and Carroll Gap, qualitative assessment on stream-aquifer interactions was not made. Probably some of the reaches downstream of Tamworth are gaining, specifically near the Carroll Gap area (Berhane and Serov, 2004).

In addition, an important hydrological characteristic for water resources management is the persistence of flow. For instance, the Peel River at Bowling Alley Point (419045) ceased to flow on many occasions during droughts (WRC, 1970). The maximum-recorded period of zero flow was 69 consecutive days, which occurred over March and April, 1920 (WRC, 1970), prior to irrigation development.

The upland areas of the Peel Valley, including the Cockburn valley are similar to the Macdonald sub-catchment. The FDC for the Macdonald River at Woolbrook (419010) showed a moderate degree of persistence during long dry periods (Berhane, 2006). This would appear to result from the inability of the groundwater storage in the valley to sustain low flows in streams during extended periods without significant rainfall Bradd (1994).

2.5.7 Mooki and Mid-Namoi Sub-Catchments

Bradd (1994) provided preliminary information on the nature of recharge and discharge of groundwater and aquifer connectivity based on an isotope and geochemical survey in the Mooki River Valley. The first order creeks draining the upland catchment (Yarrimanbah, Taylors, Mcdonalds, Millers, Little Jacks and Big Jacks Creek) were mapped as dry ephemeral creeks (Bradd, 1994). These lower order streams join to form the Mooki River which is basically an ephemeral river and makes little contribution to the flow in the Namoi River, except during high flow and flood events (Ivkovic, 2009b).

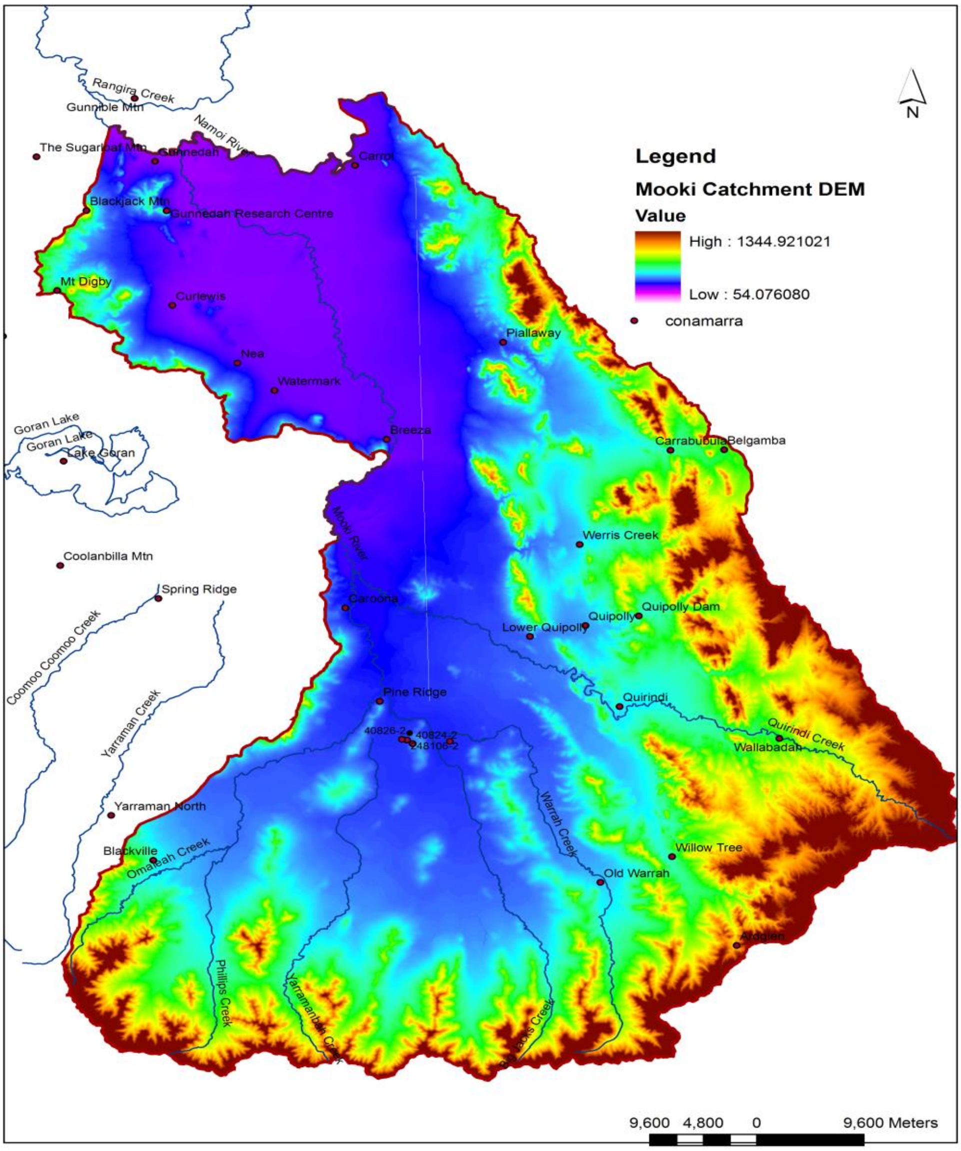


Figure 2-15. Dry ephemeral creeks systems in the Mooki sub-catchment

In contrast to Bradd's work, the Mooki River upstream of Breeza was mapped as a variably connected gaining losing reach. Upland tributaries of the Mooki were mapped as gaining baseflow from the groundwater system (Brodie et al., 2007b). From geomorphological and hydrological point of view, the lower order streams of the Namoi Valley are predominantly losing streams rather than gaining. As Ivkovic (2009a) mapping methodology at a regional scale disregarded the position of water table in relation to channel bed, the mapping technique is questionable. However, localised groundwater discharge from perched systems may occur.

Brodie et al. (2007a), used a frequency analysis approach to infer the hydraulic connectivity between streams and aquifers, by comparing daily percentiles of streamflow and rainfall. Three Australian streams were examined: a dominantly gaining stream (Wilsons River, NSW), a dominantly gaining stream modified by significant water extraction (Ovens River, Victoria) and a dominantly losing stream (Mooki River, NSW).

The streamflow record for the Mooki River at Breeza is characterised by significant periods of no-flow conditions. Recessions following significant rainfall tend to be steep and short-lived in contrast to the persistence of recessions evident in a high-baseflow stream, such as the Wilsons River. The recession steepness and large number of no-flow days suggests that stream loss to the shallow alluvial aquifer is a significant process (Brodie et al., 2007b; Pilgrim et al., 1988).

Zone 4 (an area between Keepit Dam and Gin's Leap in Figure 2-13) was included in the review because of its large size and the largest number of aquifer access licence holders in the Upper Namoi Valley. In addition, groundwater levels in the riparian zones of Zone 4 are regularly recharged by the regulated reaches of the Namoi River, and are considered to be more stable in comparison to non-regulated riparian zones (for example in Zone 3). However, the extent of the localised cone of depression is a function of pumping rate.

Historically, at a reach scale, there is disagreement whether Zone 4 is gaining or losing. For instance, some of the early groundwater modelling reports (Brodie et al., 2007b) divided Zone 4 into two sections: losing reaches of upstream of Barbers Lagoon and a gaining section downstream of Barbers Lagoon. Downstream of Barbers Lagoon the groundwater levels are generally above the stream bed system, and therefore the aquifer is feeding the stream system. Although the role of constrictions on groundwater flow is not clear, some of the reports, Lawson and Treloar (1989) speculated that the constriction in the alluvium near Boggabri, in conjunction with the Coxs Creek inflow are considered to be the cause of the high water table levels just upstream of Boggabri. The combination of high tributary inflow from Maules Creek and a general water contraction near Narrabri are considered to be the cause of the generally high water table levels between Boggabri and Narrabri, the gaining stream section of the valley.

In contrast, a recent hydrogeological investigation carried out in Zone 4, the Namoi River in this section of the river reach has been mapped as a disconnected losing stream[7]

As part of disconnected rivers project, Allen (2009) conducted resistivity survey of the sediments below the river bed for the reach between Mollee and Gunidgera Weirs. The systems have been mapped either connected or disconnected (Figure 2-16). Resistivity survey of streambed material can be used to infer whether a river system is connected or disconnected.

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[7] http://www.namoi.cma.nsw.gov.au/78829.html?4: KLC Environmental Pty Ltd, 2010).

Based on this brief overview of the Namoi River, the inland river systems of NSW are drained by ephemeral and intermittent river systems, which are predominantly losing at a catchment scale. Given contrasts in climate and hydro-geomorphological features, are the coastal river systems different from the inland river systems?

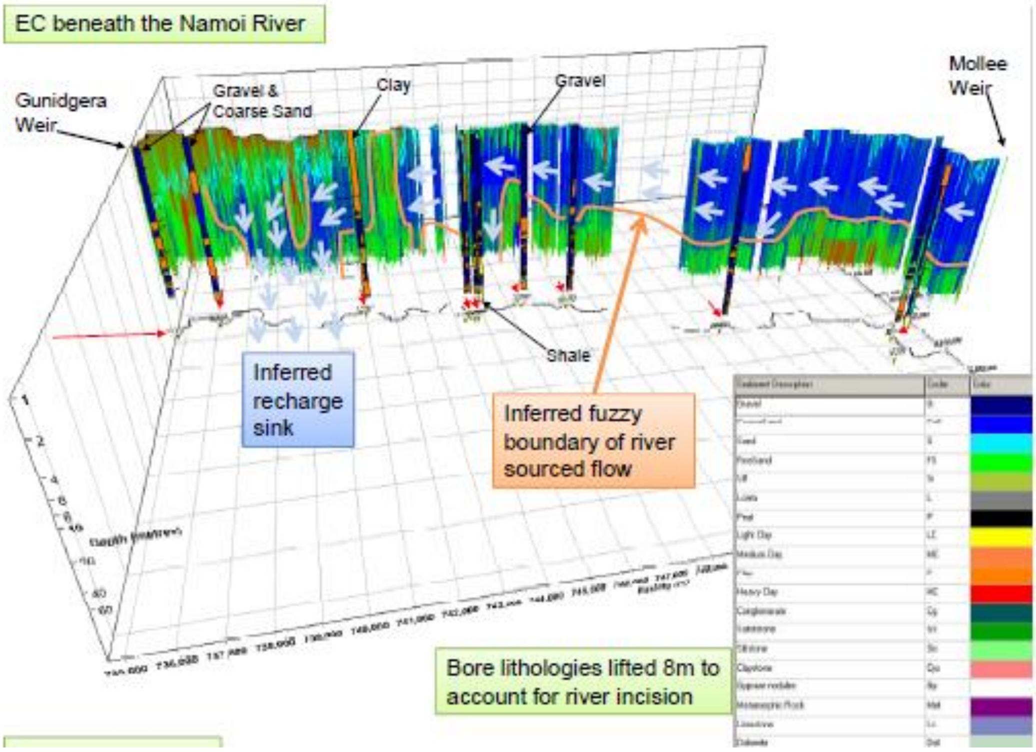


Figure 2-16. The Lower Namoi River is a losing reach between Guidgera and Mollee Weirs

For coastal catchments, Pilgrim et al. (1988) reported that TL constitutes a big proportion of the water balance equation. Consequently, TL as hydrological process is not only restricted to inland catchments and but also occurs in the coastal environments with high rainfall regime. The main conclusions of Pilgrim's work in the context of this review are: 1) TL occurs not only in lower rainfall regions, such as the Upper Hunter Valley with a median annual rainfall of 650 mm, but also in wetter districts, such as the Clarence valley with annual rainfalls between 1000 and 1800 mm; 2) the rate of TL is greatest in the headwaters of a catchment, while the total loss is greater in the large basins; 3) TL is important in small headwater channels in eastern NSW as well as in the large downstream channels.

Furthermore, Ward (1984) highlighted some aspects of river flow in northern NSW by using the FDC. All the western rivers and most of the coastal rivers have steeply sloping FDCs, which denote highly variable flows with a large quickflow component. Only the Nymboida and the Hastings show signs of having a flattened lower end to their FDCs, indicative of significant amounts of perennial storage within the drainage basin Ward (1984).

2.6 Discussion of Misconceptions related to Ephemeral River Systems

After reviewing the international literature and Australian literature, reports on TL in Australia are biased toward the more arid interior of the Lake Eyre Basin, where high TL and low water availability has created more interest in the past three decades. In this arid region of the Australian continent TL can reach up to 100% of the total runoff. However, after the introduction of water trade and other water reform agenda introduced after the millennium drought, our understanding of TL in semi-arid regions of Australia is evolving.

An understanding the basic components of TL is required to formulate proper water resources management in dryland river systems. Despite the four parts of TL (billabong/floodplain storage, infiltration, evapo-transpiration and bank storage) representing a loss component of the water balance equation, due to inconsistency persisting in literature on TL and other important hydrological processes, such as baseflow, this discussion focuses on misconceptions related to ephemeral river systems in the context of arid and semi-arid environments of Australia.

2.6.1 Diffuse vs. Focused Recharge

A quantification of recharge is a pre-requisite for effective water resources and ecosystem management. Recharge is generally considered to occur in topographic highs and discharge in topographic lows in humid regions, whereas in arid alluvial-valley regions recharge is usually focused in topographic lows such as channels of ephemeral streams (Ward, 1984).

Recharge occurs through diffuse and focused mechanisms. Diffuse recharge is distributed over a large area in response to precipitation infiltrating the soil surface and percolating through the unsaturated zone to the water table (Berhane, 2006). Focused recharge is the movement of water from surface water bodies, such as streams, canals, or lakes, to an underlying aquifer. Generally, diffusive recharge dominates in humid environments. As the degree of aridity increases, the importance of focused recharge increases (Healy, 2010). From case studies in semi-arid regions of the US, some investigators reported that surface groundwater interactions in semi-arid regions take place in relatively small and distinct parts of the landscape, and suggest that episodic or variable interactions are the norm; recharge is non-existent or negligible over much of the semi-arid landscape (Lerner et al., 1990a).

Although estimation of recharge in disconnected groundwater systems involves a high degree of subjectivity, recent research work carried out in the Lower Namoi Valley (Newman et al., 2006) suggested that the magnitude of recharge is insignificant. The findings of this study are similar to the work carried out in the Western US (Figure 2-17), which challenges previously accepted estimates of widespread, low rates of deep drainage and groundwater recharge in arid and semi-arid regions (Timms et al., 2011). For example, based on chloride mass balance survey, insignificant recharge has occurred since the Pleistocene over much of the non-mountainous, inter-drainage area of the American southwest (Seyfried et al., 2005). A follow up research work using chloride and other tracers has supported the idea of little to no recharge over large areas in the southwest of the US (Duffy, 2004; Phillips, 1994).

In the Kalahari, a semi-arid region of Southern Africa, annual rainfall varies from 250 to 550 mm, however high infiltration rates in the sandy deposits, high retention deposit during the wet season and subsequent high evapotranspiration rates during dry season, limit percolation or recharge rate (Scanlon, 2004; Walvoord, 2002). Because of the complexity of the hydrological processes taking place in the unsaturated zone and hydroecological factors a lively debate has continued for almost a century on the question of whether the Kalahari aquifers are being recharged at all under present climatic conditions (de Vries, 2002b).

In the context of wadis of North Africa and the Middle East, focused recharge is considered to be the dominant form. Water balance studies carried out for the Tihama coastal plain bordering the Red Sea in Yemen indicate that around 60 percent of groundwater recharge is derived from wadi flows (de Vries, 2002b) reported that infiltration of wadi flows provide the major source of recharge to the aquifers of both the Abyan and Tuban deltas in Yemen. Apart from a quantity of subsurface inflow, wadi flows provide the only source of replenishment for the aquifers. Other recharge components are merely infiltration of diverted spate flows or recycling of abstracted groundwater. A water balance study carried out for Wadi Turban indicated that approximately 48% of the surface inflow recharged the aquifer by infiltrating from wadi beds (DHV, 1988; Komex, 2002).

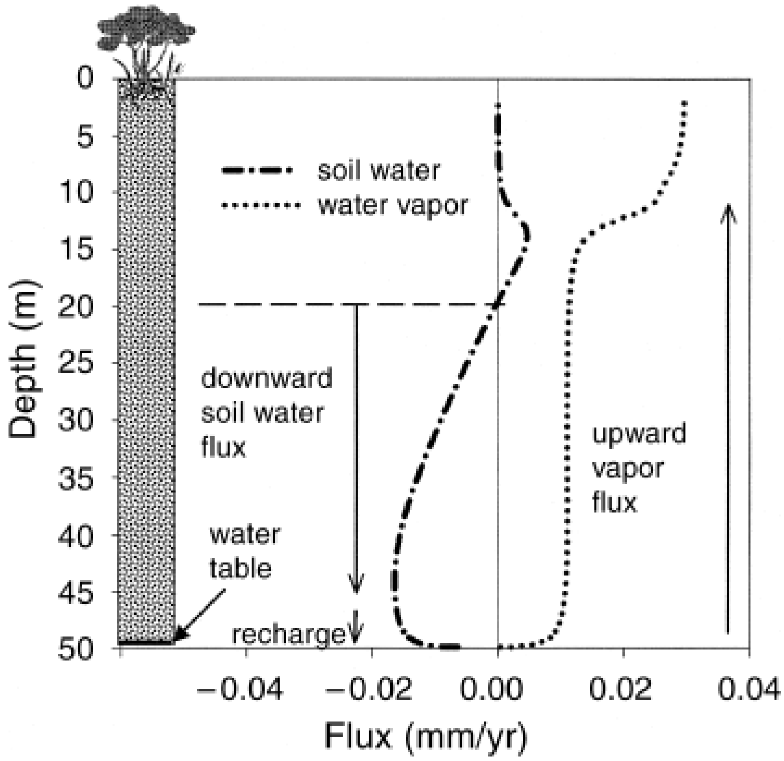


Figure 2-17. Model-predicted soil water (liquid) and water vapour fluxes for depths below the root zone 10 kyr subsequent to the establishment of xeric vegetation. Negative values indicate downward fluxes (after Seyfried, 2005).

Hydrologic processes in disconnected groundwater systems of water-limited environments unfold over longer timescales than those in surface and near-surface soils (Komex, 2002). Quantification of recharge thus is beyond simplistic models and would require deep vertical profiles of water content and water potential (to ascertain if gradients favoured upward versus downward water movement) along with chloride profiles (to quantify recharge by the chloride mass balance method (Newman et al., 2006).

As part of a project to assess the impact of pumping on groundwater quality, the Office of Water-NSW conducted a detailed groundwater quality survey in Zone 3 of the Upper Namoi Valley in early 2000. The areal distribution of salt in the shallow groundwater system is presented in Figure 2-18 which shows high level of dissolved solids reaching up to 10,000 mg/l and groundwater ages ranges from 10,000 to 20,000 years (Newman et al., 2006). The high level of groundwater salinities is not restricted to Zone 3. This is a common phenomena in all groundwater management zones in the Namoi Valley. In fact, in the vicinity of Goran Lake area, salinity levels can reach up to 30, 000 mg/l (Bradd, 1994).

Similarly, in the Lower Balone area, recharge rates based on chloride mass balance ranged from 0.05 mm/yr for the highly saline upper alluvial and meta-sediment formation waters to 5 mm/yr for the fresh groundwaters in the upper alluvial and middle confining bed (Lavitt, 1999).

Radiocarbon ages for the lower alluvial aquifer tend to be much older (>5,000 - >25,000 years) than the upper alluvial aquifer (<2,500 year) suggesting that vertical leakage occurs at a very slow rate. Fresh, relatively young (<50 years) groundwater may be restricted to near the rivers suggesting flood recharge is limited in lateral extent via surface tributaries, but further work would be required to confirm this conclusion (Herczeg, 2004). There is evidence for vertical exchange of the groundwaters amongst the various aquifer systems (indicated by chemistry and stable isotope data), however this occurs on time scale of 1000 to 10,000 years as indicated by ‘old’ 14C groundwater ages at depths >20 m below the water table (Herczeg, 2004).

Given the old age of groundwaters and high salt contents in some areas of the Northern MDB, how does distributed groundwater recharge takes place in this type of environment? What is the magnitude of groundwater recharge in the current geological time scale?

In the Northern MDB catchments, the superficial aquitards host saline groundwater of different chemical compositions and ages. Thus, the main factors controlling the salinity of groundwater in the underlying aquifer units are the thickness of the clayey aquitard layers, episodic infiltration rates and flushing rates of pore water saline water entrapped in the aquitards layers.

In summary, there is high uncertainty of recharge estimates below 5 mm and possibly as high as 20 mm. Due to the fact that the upward movement of water vapour through the profile to replace the evaporation losses from the top 2 m of the soil is usually not considered (FAO, 2010). Thus, EC or Cl concentrations can be used as a surrogate indicator of rate of flushing or groundwater recharge rate.

Due to constraints in resources and time, most of the groundwater recharge estimates in the Northern MDB are based on mechanistic models. Unfortunately, these models push water in any system, and it is up to the analyst to comprehend the limitations of models for estimation of recharge in water limited, arid and semi-arid environments as well as partition of gross recharge into focused and distributed components.

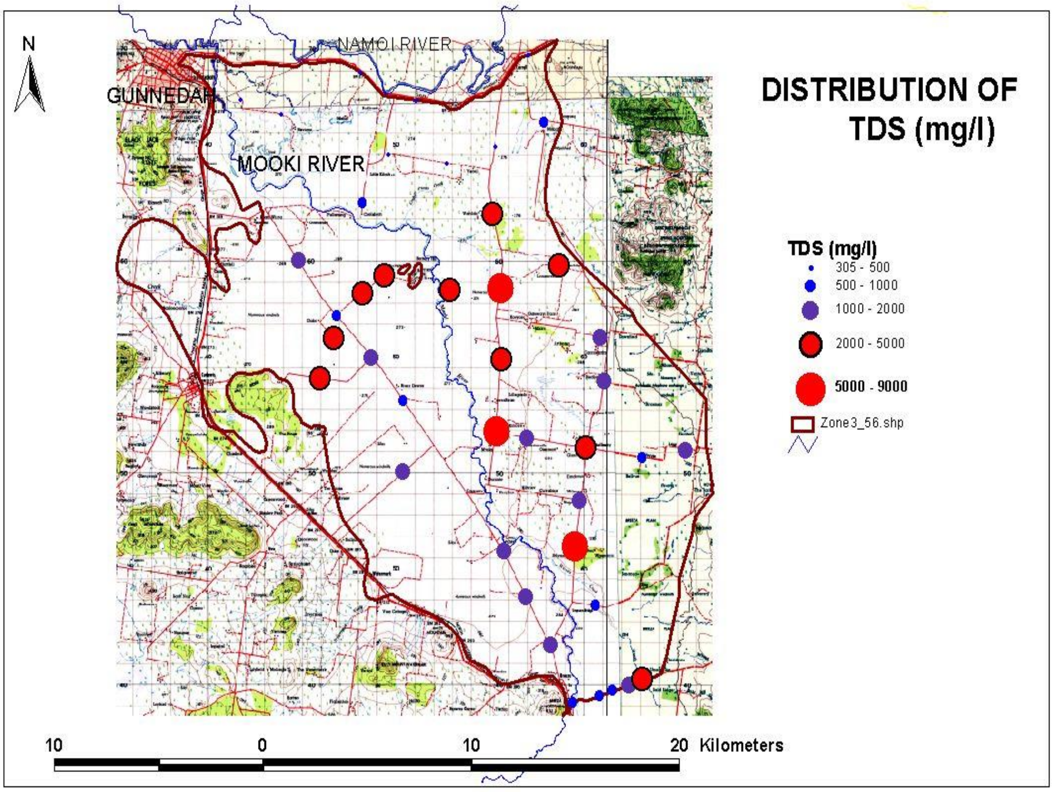


Figure 2-18. Total dissolved solids in shallow groundwater system of Zone 3, the Upper Namoi Valley (Unpublished report, Berhane, 2004)

2.6.2 Transmission Losses: Variations in Space and Time

Braaten and Gates (2003a) reported the relationship between river-aquifer connection and geomorphological positions in the MDB at a basin scale (Figure 2-19). According to this study upland streams located in fractured rocks have been mapped as gaining systems. Narrow alluvial valleys have been mapped as gaining/losing systems, while a mid-slope section of the cross-section, representative of extensive alluvial plains where major groundwater development takes place, which has been mapped as a losing disconnected system. The lower reaches of the cross-section were mapped as variably gaining and losing.

At a smaller scale, water exchanges in the streambed occur simultaneously on various scales (Herczeg, 2004) but a basic unit of exchange occurs across riffle-pool sequences. On this scale, surface water enters the subsurface sediments at the upstream end of a riffle (a shallow, fast flowing section of a stream) and returns to the stream channel at the downstream end (Braaten and Gates, 2003a). Riffle-scale exchange flows are of particular significance to the stream ecosystem as they are prevalent and likely account for more interaction (Chapter 5 of this thesis). However, the volumes of water involved in the exchanges are typically insignificant as a fraction of the stream flow. The significance of hyporheic exchange is in the movement of solutes from the stream into the subsurface (Hendricks and White, 1991; Valett, 1994).

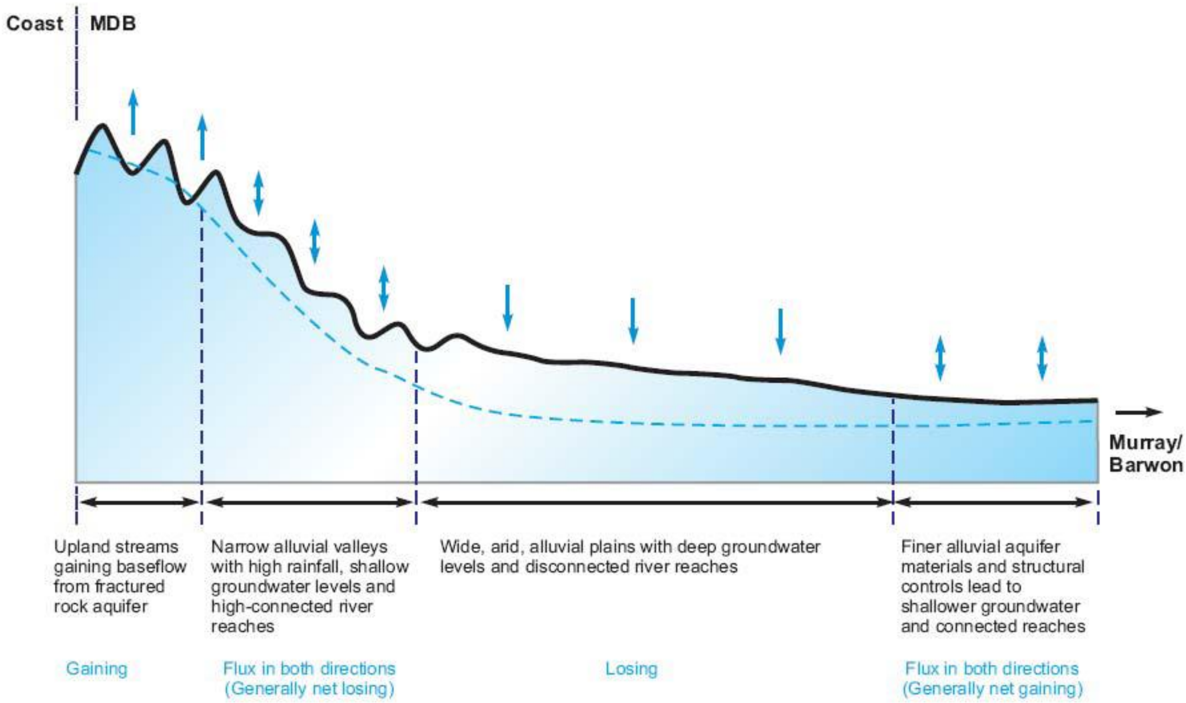


Figure 2-19. Transect of Murray-Darling Basin showing relationship between river–aquifer connection and landscape position. Blue arrows indicate the direction of river–aquifer flux

The relative importance of factors that cause TL varies among arid catchments depend on characteristics, such as the size of the catchment, stream gradient, stream/floodplain sediments and seasonal timing of flow (Costelloe et al., 2003), and climate. In small to medium sized catchments with low to moderate gradients in Arizona (Braaten and Gates, 2003b); India (Walters, 1990) and South Africa (Sharma, 1994), infiltration to the near-surface channel aquifer was the major cause of TL (Sharma and Murphy, 1994).

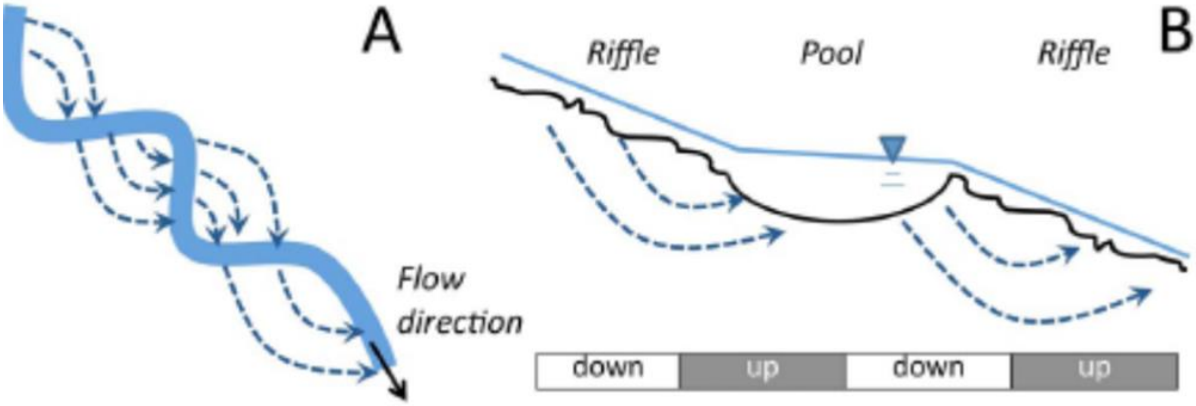


Figure 2-20. Idealized representation of hyporheic exchange in (a) plan view (lateral exchange) and (b) vertical cross-section (vertical exchange) (after Hester and Gooseff, 2010)

In large catchments such as Coopers Creek in Western Queensland and the Orange River in South Africa evapotranspiration was considered as the principal cause of TL. In addition, the size and duration of storm matters. Runoff generated during relatively small storm events has to satisfy in-channel waterholes before producing flow in downward direction (Hughes and Sami, 1992).

In summary, TL is complex, non-linear hydrological phenomena determined by evapotranspiration, infiltration and water required to fill waterholes and channel storages as river water moves down a stream. Most of the Australian river systems exhibit TL of different magnitudes depending on the climate, geology and geomorphology (Morin et al., 2009).

2.7 Selection of method: heat as an environmental tracer

2.7.1 Heat tracing technique

Based on Flint et al. (2002), the advantages of estimating infiltration by heat tracing technique are: 1) temperature is easy to deploy and monitor with accuracy; 2) the method is relatively in-expensive; 3) the relation of thermal conductivity to water content is linear, and temperature is sensitive to fracture and matrix flow. However, to be able to detect significant changes in temperature profiles, fluxes of 10 to 20 mm/year are required (Flint et al., 2002; Healy, 2010). In addition, the traditional instrumentation can only deal with small-scale losses or provides a point scale assessment of infiltration rate. However, the emerging fiber optic technology can be used for regional scale investigations. Thus, based on cost factors and suitability of the hydrothermic environment in the study area, heat as an environmental tracer was selected for this field investigation in selected catchments of the Namoi Valley and Central West Catchments. In addition, the method is an ideal tool for communication or reporting to the landholders on complex hydrological processes in relation to recharge and discharge. The role of communication should not be underestimated in catchment as the Namoi Valley, where stakeholders are bit suspicious of different technical messages originating from different groups.

2.8 Knowledge Gaps and Opportunities

2.8.1 Knowledge gaps

2.8.1.1 Scouring of riverbed sediments and infiltration

The riverbed of Coopers Creek and other river systems in the LEB consist of thick clay layer, which are considered to have been deposited during Holocene and possible late Pleistocene (Fagan and Nanson, 2004); Scanlon (2004). In some parts of the channel, the riverbed is underlain by coarse to medium quartz sands deposited, when Cooper Creek sustained much higher discharges during the Quaternary (Larsen, 2012). During episodic flood events, the channel can have a sufficient stream power to scour the clogging layer and enhance surface water infiltration into the underlying unsaturated zone. Once the floodwaters cease, the mud carried in suspension settles to the base of the channel once more, thus sealing the channel and isolating it from the water table (Nanson et al., 2008). A scouring induced infiltration mechanism has been observed to occur in the inland river systems of the Namoi Valley, where the groundwater quality is fresher than the GAB shallow groundwater system. Thus, further research work is required to quantify the extent of fresh water lens that may occur in the LEB post flood events. As a start, some aspects of clogging and scouring processes that could influence magnitudes of TL are discussed in Chapter 4 of this thesis.

2.8.2 Springs in arid and semi–arid environments

Probably early human and ecosystem survival in arid environments of Australia would not have been feasible without springs, waterholes and other water storage structures. The artesian discharge from the Great Artesian Basin feeds spring wetlands that can be distinguished from other wetlands in dry landscapes because of their permanent water source[8]

[8] http://www.environment.gov.au/soe/2006/publications/emerging/wetlands/

Since groundwater development started in the GAB region, groundwater pressures have declined in a number of regions in the basin leading to reduced flow in many artesian bores and springs. Although some of the bores have been capped during the last two decades, the resource is being depleted through uncontrolled artesian discharge. This has implications for water users, including ecosystems that are dependent on the mound spring water supply. In the eastern GAB recharge areas (of which a small portion lies within the LEB) ‘overflow’ springs provide important ‘baseflow’ to rivers (Larsen, 2012). At this stage, the extent of human induced impacts on springs and wetlands in the GAB has not been assessed in detail.

In Chapter 3 of this thesis the hydrology of disconnected river systems is discussed. However, during literature review, it became clear that there is a knowledge gap in terms of quantifying stream depletion of losing streams. Although (McMahon et al., 2005) discuss some of the issues related to disconnected river systems and the factors that influence the dynamics of disconnection; there is still lack of knowledge on environmental flow management of disconnected river systems.

2.8.3 Opportunities

2.8.3.1 Use of fiber optics technology

Temperature traditionally has been reported to be an excellent environmental tracer to study surface groundwater interactions (Brunner et al., 2011). Natural variations in stream water temperature patterns are used to assess the interaction of river water with shallow groundwater. The temperatures of surface water bodies are variable and change in response to seasonal and meteorological changes, in contrast groundwater temperatures are relatively stable year round. Groundwater discharge into surface water can be visualised by negative anomaly (cold signature) in summer and positive anomaly (warm signature) in winter.

During the last decade, Fiber-optic distributed temperature sensing (FO DTS) technology supplements conventional discrete-point and remote sensing technologies for temperature measurement (Stonestorm, 2003). FO DTS utilizes laser light to measure temperature along the entire length of standard telecommunications optical fibers. The technology can measure temperature every meter over FO cables up to 30 kilometres (km) long (Day-Lewis and Lane, 2006). Commercially available systems can measure fiber temperature as often as 4 times per minute, with thermal precision ranging from 0.1 to 0.01 °C depending on measurement integration time (Day-Lewis and Lane, 2006). This configuration is an ideal setup for delineation reaches with focused recharge. However, for delineation of losing reaches, temperature should be measured in bores or at different depths under streambed or close to the bank of a river. For instance, (Day-Lewis and Lane, 2006) studied near streambed surface water interactions by applying distributed temperature sensing to an FO DTS wrapped around a tube that is installed in the river bed. The authors concluded that the FO DTS technology enables to sample high quality temperature data for quantification of stream leakage. Thus, DTS is an attractive method to obtain long profiles of temperatures that vary on time scales Vogt et al. (2010).

2.8.3.2 Use of Remote Sensing

Evapotranspiration (ET) constitutes a large component of TL and the hydrologic cycle and is considered as the important parameter in the water budget in semi-arid regions of the MDB. Estimation of ET at a catchment and finer scales is important activity in improving the management of the limited fresh water resources in the MDB. Today, estimation of ET using remotely sensed data is an emerging issue in water limited environments of the world (Selker, 2006); therefore, this technique should be refined and applied on a regular basis.

Several research groups are now developing tools to manipulate satellite data more easily. An example of this is that recently researchers at the Center for Research in Water Resources at the University of Texas – Austin have developed a toolbox-MODIS Toolbox, which facilitates the importation of MODIS products (evapotranspiration, land surface temperature, normalized difference vegetation index (NDVI), and enhanced vegetation index (EVI)) into the ArcGIS environment.

2.8.3.3 Groundwater Age and Age Dating

Groundwater age and age dating form a critical link between different facets of hydrology (Bethke and Johnson, 2008). Our understanding of groundwater age has evolved over the last two decades. Currently a groundwater sample is seen, not as water that recharged the flow regime at a point in the past, but as a mixture of waters that have resided in the subsurface for varying lengths of time (Bethke and Johnson, 2008).

Estimation of groundwater recharge is important activity for management of water resources. Thus groundwater age dating may in the future assist in our understanding the rate of fluxes in a complex hydrogeological system such as in the Namoi Valley (Lavitt, 1999). Bethke and Johnson (2008) have demonstrated in an aquifer/aquitard interface, that the mean age in the aquifer is not controlled by the effects of aquifer-aquitard mixing rate but, by the ratio of fluid volume in aquitards to aquifers (the so-called “paradox of groundwater age”, where at low mixing rates, very old water is supplied to the aquifer, but the water in the aquitard remains old, while a high mixing rates, less-old water is supplied to the aquifer, because younger water is moving into the aquitard).

2.8.3.4 Gravity Survey

GRACE has been proven reliable and offers a great potential for water storage budget closure on basin to regional scale (Anderson et al., 2012). In the future, precision gravity surveys may possibly be used to monitor recharge rates or changes in groundwater storage.

At reach and small scale investigations, microgravity determinations of sub-surface storages would probably be an interesting research tool in the near future for a better understanding of flow interchanges in ephemeral streams (Goodrich et al., 2004).

At a regional scale, new research using data from NASA’s gravity-sensing GRACE satellites show a dramatic decline in the volumes of groundwater reserves in the Tigris and Euphrates river basins. Data gathered between 2003 and 2009 suggest that the Tigris and Euphrates river basins lost about ~144 km^3 of total stored freshwater (Voss et al., 2013). In the case of the MDB, a total water loss ~140 ± 54 km^3 was reported between August 2002 and December 2006 (Voss et al., 2013), this is an accumulated reduction of ~130 mm equivalent water depth across the basin.

Recently, for the Border Rivers region, recharge has been estimated at different spatial and time scales, where GRACE satellite data was used to capture changes in total water storage at a basin scale (Berhane et al., 2015). However, downscaling of GRACE data to the scale of interest remains a challenging exercise.

Chapter 3 Visualisation of Groundwater Mounding in arid and semi-arid environments

"As a component of the water balance equation, transmission losses are the dominant processes in arid and semi-arid regions of NSW; occurrence of groundwater recharge is an episodic event"

This thesis, 2015

Abstract

Prior to the application of heat as an environmental tracer, an insight into hydrological processes of ephemeral and intermittent streams was developed by simulating hypothetical head distributions using the VS2DT package. The results from the VS2DT simulations suggest that the relationship between seepage and groundwater level is highly variable and non-linear for a disconnected river with a shallow water table setting. Although highly variable, depending on the stream stage and groundwater heads, seepage flux reaches a constant value below a certain critical depth in disconnected river systems, for an isotropic and homogenous aquifer system. This critical depth of water table, where river seepage flux attains a constant value depends on channel geometric configuration and hydraulic conductivity of the streambed and underlying aquifers.

3.1 Introduction

The inland rivers systems in NSW are characterized by high variability of flow and extended droughts and dry spells. Based on historical stream flow data, unregulated streams in the Central West and Namoi Catchment can be classified either as intermittent or ephemeral, depending on their positions in the landscape. Due to the climate signature of the inland river systems, perennial streams with continuous stream flow are not prevalent in the region, even on the Tablelands, where rainfall ranges from 650 to 1200 mm/year (Chapter 2).

At the catchment scale, Thoms et al. (1999) reported that the long-term average annual runoff for the Namoi River is 770,000 ML at Gunnedah, which represents 6% of the average annual catchment rainfall. Most of the runoff is generated in the headwaters of the valley (Cockburn, Peel and MacDonald catchments), about 90% of the total runoff comes from 40% of the catchment area Thoms et al. (1999). In the upland catchments, annual flows increase with increasing catchment area, but downstream of Gunnedah annual flows decrease due to increasing evapotranspiration, infiltration and water use for irrigation (Thoms et al., 1999). In ephemeral and intermittent streams of the northern Murray Darling Basin (MDB) catchments, the standard deviation exceeds the mean of the flow; thus, the term average should not be taken at a face value.

This chapter was compiled as background to Chapter 4, where basic hydrological processes of connected and disconnected river systems are visualized using hypothetical 2D cross-sectional models, while in Chapter 4, results from the application of heat as an environmental tracer are presented and analysed. The first section on visualisation was implemented using a VS2DT package and is expected to provide some insights on hydrological processes below the streambed. As the unsaturated zone is a buffer between the channel (surface water bodies) and groundwater system, accurate calculation of the magnitude and occurrence of recharge is only possible by comprehending the physical processes in the unsaturated zone.

Therefore, the main objectives of this study are to: 1) visualise groundwater mounding in connected and disconnected groundwater systems using a physical-based model, and gain insight in some hydrological processes of these systems; and 2) develop relationships between water table and stream seepage rate for a given simple stream channel geometry and other hydraulic parameters.

3.1.1 Concepts

For consistency purpose, some basic definitions on hydrological terms used in this chapter are provided below.

- **Armouring (of sediments):** Formation of an erosion resistant layer of relatively large particles resulting from the removal of finer particles.
- **Base flow:** Discharge which enters a stream channel mainly from groundwater (perched, local, intermediate and regional systems), during long periods when no precipitation or snowmelt occurs.
- **Clogging layer:** a thin soil on streambeds, which consist of fine sediments, impeding infiltration. Clogging is caused by physical, biological, and chemical processes (Baveye et al., 1998). Physical processes comprise the accumulation of inorganic and organic suspended solids in the recharge water, such as clay and silt particles, algae cells and microorganism cells and fragments (Bouwer, 1978).
- **Hydrostatic condition:** a point on the vertical profile, where groundwater head equals the stream head.
- **Incipient disconnection:** a point where the water table disconnects from the surface water.
- **Maximum incipient seepage:** a point at which maximum stream leakage occurs, which is a transition from head variable boundary into constant boundary condition in a model.
- **Diffusivity:** ratio of the transmissivity to the storage properties of an aquifer: T/S, K/S, or T/Ss.
- **Stream depletion:** stream depletion implies decreasing river stage induced by pumping from a nearby production bore.

- **Hyporheic zone:** an ecotone between stream water and groundwater environments, combining not only biogeochemical but also physical characteristics of both sources. The hyporheic zone provides an ideal habitat for a wide array of microbes and invertebrates. In arid and semi-arid regions, the term streambed zone may be an appropriate terminology.

3.2 Background theory

In the hydrological literature the definition of ephemeral and intermittent streams is a bit unclear. Ephemeral channels are distinguished from intermittent and perennial streams based on duration of flow. Perennial streams flow continuously except in extreme drought, and intermittent streams flow during the wet season (Fritz, 2008). Ephemeral channels commonly lack defined banks and generally have large amounts of organic matter in the channel bed (Hansen, 2001). Based on the definition of Hedman and Osterkamp (1982) perennial streams are considered as having measurable discharge 80% of the time, intermittent 10-80% of the time, and ephemeral <10% of the time.

Similarly, there is also a lack of common ground on the definition of disconnected/perched river systems. For example, Bouwer and Maddock (1997) provided a criterion for disconnection, when the depth of the water table below the stream channel exceeds twice the stream width. In contrast, work carried out by (Brunner, 2009a; Jackson, 2005; Osman and Bruen, 2002; Rushton, 2007) suggested that the critical water table depth, where a system disconnects depends on multiple parameters, which include the geometric configuration of the streambed, hydraulic conductivities of the clogging layer and aquifer, and depth to groundwater. According to Bouwer and Maddock (1997) a stream is considered to be disconnected if an unsaturated zone between the river and water table exists and the total infiltration flux across the stream width attains a constant value.

While extensive fieldwork can provide insights into stream aquifer connections, i.e., Lamontagne et al. (2014), this is often expensive and delivers a snapshot of behaviour. Alternatively, some research has applied simulation modelling to determined aquifer connection, i.e., Brunner (2009a). Simulations using coupled unsaturated and saturated models are a challenging exercise because lack of knowledge on unsaturated hydraulic conductivity and soil properties in general and these values are highly variable, depending on the moisture content of the unsaturated zone. Thus, at a regional or catchment scale, channel transmission losses are the most difficult hydrological processes to assess (Parissopolous and Wheather, 1992); (Knighton and Nanson, 1994a).

Furthermore, an important hydrological question related to environmental flow (EF) is whether stream depletion occurs in disconnected river systems. Obviously, in perennial and 'intermittent' streams there is direct linkage between groundwater pumping and stream depletion. Thus, the adverse impacts of groundwater abstraction on streamflow depletion and wetlands define a limit to the exploitable groundwater resources. For example, Sophocleous (2002) states that the sustainable yield of an aquifer must be considerably less than recharge, if adequate amounts of water are to be available to sustain both the quantity and quality of streams, springs, wetlands and groundwater dependent ecosystems. Groundwater and surface water are seen as intimately connected systems (Sophocleous, 2002; Winter, 1998; Woessner, 2000), advocated a holistic approach in management of water resources, where surface water and groundwater are considered jointly. Most of the cited technical examples by Sophocleous (2002) are based mainly on case studies from humid environments where the baseflow index

(BFI) was used to illustrate the connectivity between surface and groundwater systems. The applicability of the BFI methodology for assessments of EFs in arid and semi-arid environments could be questioned.

3.2.1 Plausible Conceptual models

Prior to providing a theoretical framework on the application of heat as an environmental tracer, a brief introduction on hydrogeomorphology and plausible conceptual model(s) based on field data and observations are provided. Conceptually, the river systems of the Namoi Valley can be grouped into three major hydrogeomorphological zones. Zone one, which includes the Liverpool Ranges and Moonobi Ranges, Warrumbugles and Mt Kaputar, predominantly consists of hard rock formations (dominantly granites with metasediments and volcanics). Due to the lack of extensive productive aquifer systems in the upland catchments, the groundwater monitoring network is sparse and little is known about the connectivity between the stream and the underlying groundwater system. However, based on geomorhological and hydrogeological factors, it is possible to infer whether the head waters are recharge or discharge areas (Toth, 2009a).

The headwaters of the Namoi Valley are drained by first to third order ephemeral creek systems, which do not carry water most of the time. By their topographic positions, the upland lower order creek systems are ‘potential’ recharge areas. There is a lack of documented information on the occurrence of permanent springs in the upland catchments of the Namoi Valley, which is therefore considered to be devoid of major permanent springs.

Given the inconsistency on the connectivity issue discussed in the Namoi literature (Berhane et al., 2008; Brodie et al., 2007a; Cook, 2012; Ivkovic, 2009b; McCallum et al., 2010) a summary on relationship between stream stage and groundwater levels is provided for the different Zones of the Namoi Valley (Figure 3-2); in this type of environments, the groundwater systems in the Namoi Valley act as sink rather than sources of discharge. For example, the Caroona area is located on the foot slopes of the Liverpool Ranges; the river system in this region appears to be disconnected even in the early 1970s, when the groundwater level was about 10 m below channel elevation (Figure 3-2). As time progressed, the level of disconnection increased due to extensive groundwater pumping that took place in the region. This area can be considered a representative for Zone I.

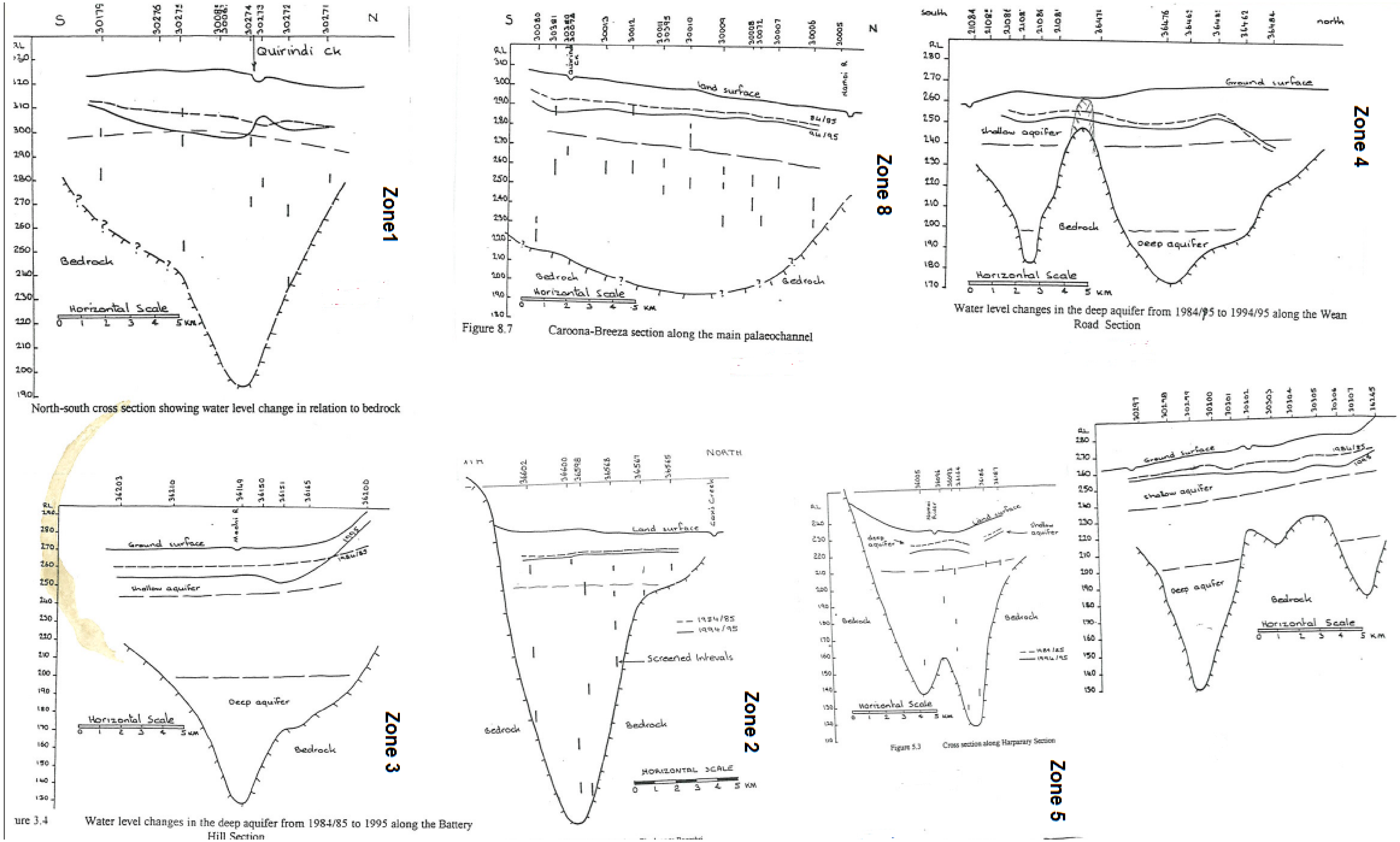


Figure 3-1 . Hydrological cross-sections in the Namoi Valley (After NSW-DLWC, 1985)

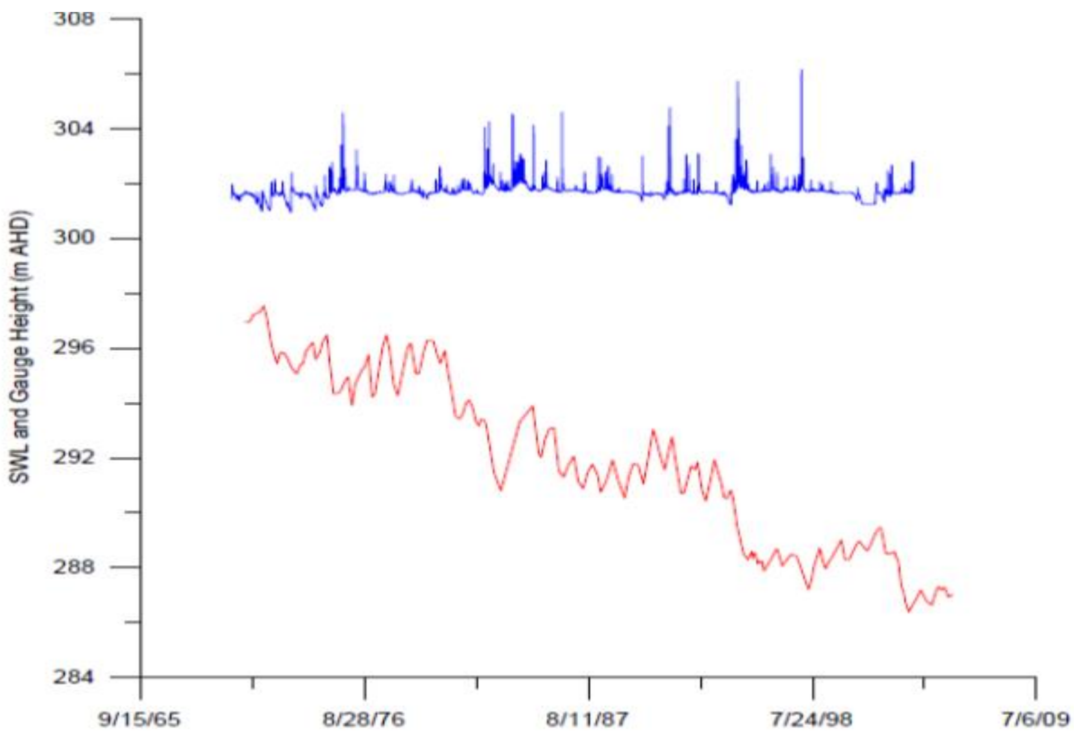


Figure 3-2. Position of water table (red) in relation to stream stage (blue) in the vicinity of the Caroona gauging station (419034).

Zone II includes areas such as the Middle to Lower Peel, the Liverpool Plains and the Pilliga area. This zone is considered to be a transition from the headwaters to the Lowland Plains, where the position of water table ranges from zero to up to ten meters below the streambed.

However, at Gunnedah gauging station the relationship between stream stage and groundwater levels is more complex and can be described as a representative of a switching hydrologic regime (Figure 3-3), the stream both loses and gains water depending on the prevailing hydrologic regime (local extraction, stream flow rates). At this particular site, the depth to water table is mainly regulated by a shallow depth to bedrock, which seems to pond water.

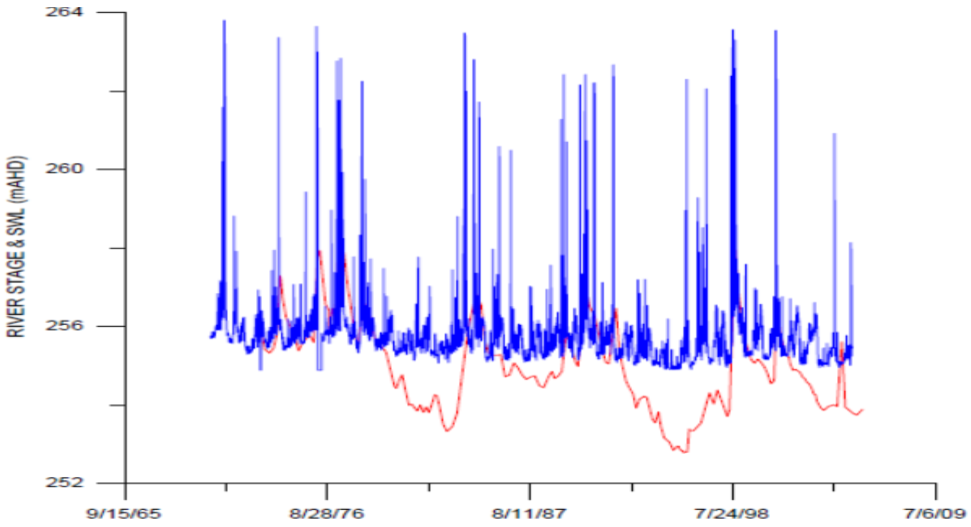


Figure 3-3. Position of the groundwater levels (red) in relation to the stream stage (blue) for the Gunnedah gauging station (419001).

Figure 3-2 illustrates the dynamic nature of groundwater levels in relation to stream stage. Based on the available hydrological information, the groundwater level is lower than stream stage, even prior to extensive irrigation practice started in the Namoi Valley in late 1960s (Lower Namoi Valley).

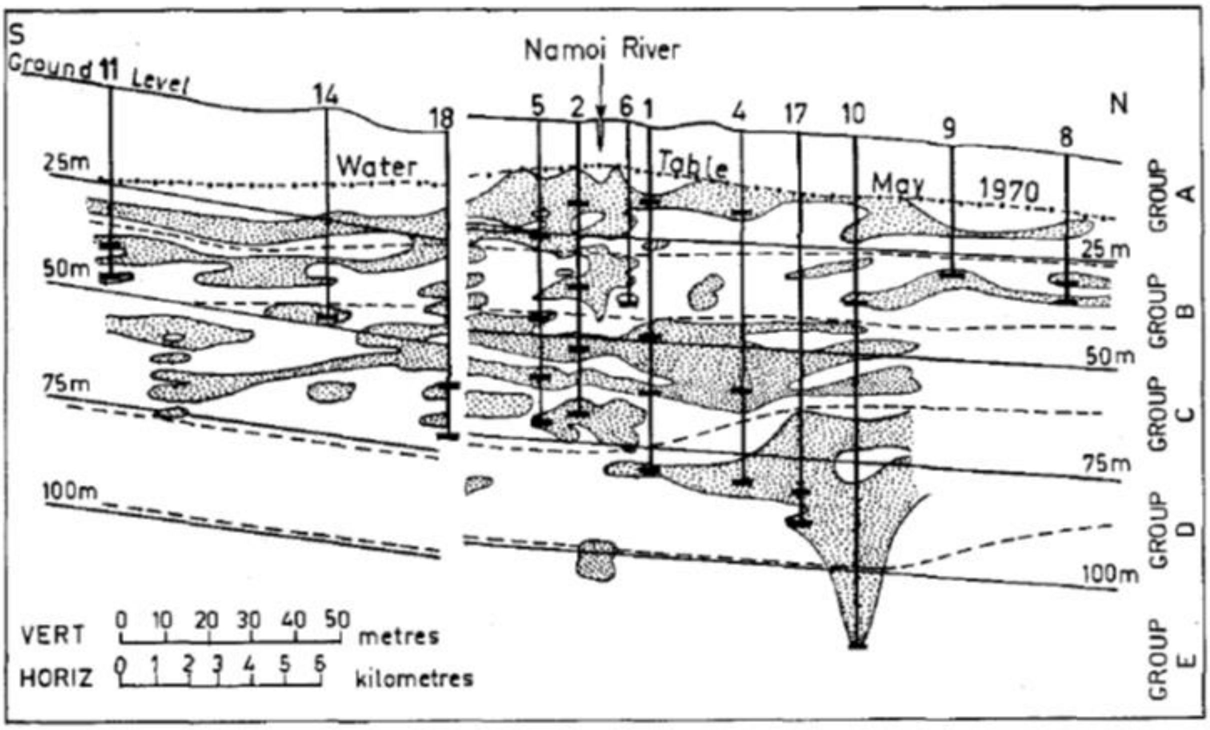


Figure 3-4. Vertical cross-section of the Gurleigh section of the Namoi Valley (after, Calf, 1978).

Figure 3-5., shows a hypothetical or idealised relationship between stream stage and groundwater head for three distinct hydrological regimes as discussed in the previous paragraph in the context of the Namoi Valley. Case A, conceptualises a hydrologic regime, where water table is higher than streambed elevation and represents a gaining stream situation. This occurs at some time in Zone 2, discussed above. While case II, represents an ‘intermittent’ stream, where groundwater head is below the stream stage. For simplicity, most groundwater models assume the groundwater system is connected and provide mathematical simulations for cases A and B (Figure 3-5.) based on the concept of leakage coefficient (Rushton, 2007). This procedure has been successfully applied to catchments and river reaches, especially in temperate and humid regions, linking distributed river and groundwater flow models (Costa et al., 2012) However, the leakage coefficient concept failed to model disconnected losing streams, because it neglects unsaturated flow through the alluvium (Brunner et al., 2010; Lamontagne et al., 2014; Rushton, 2007).

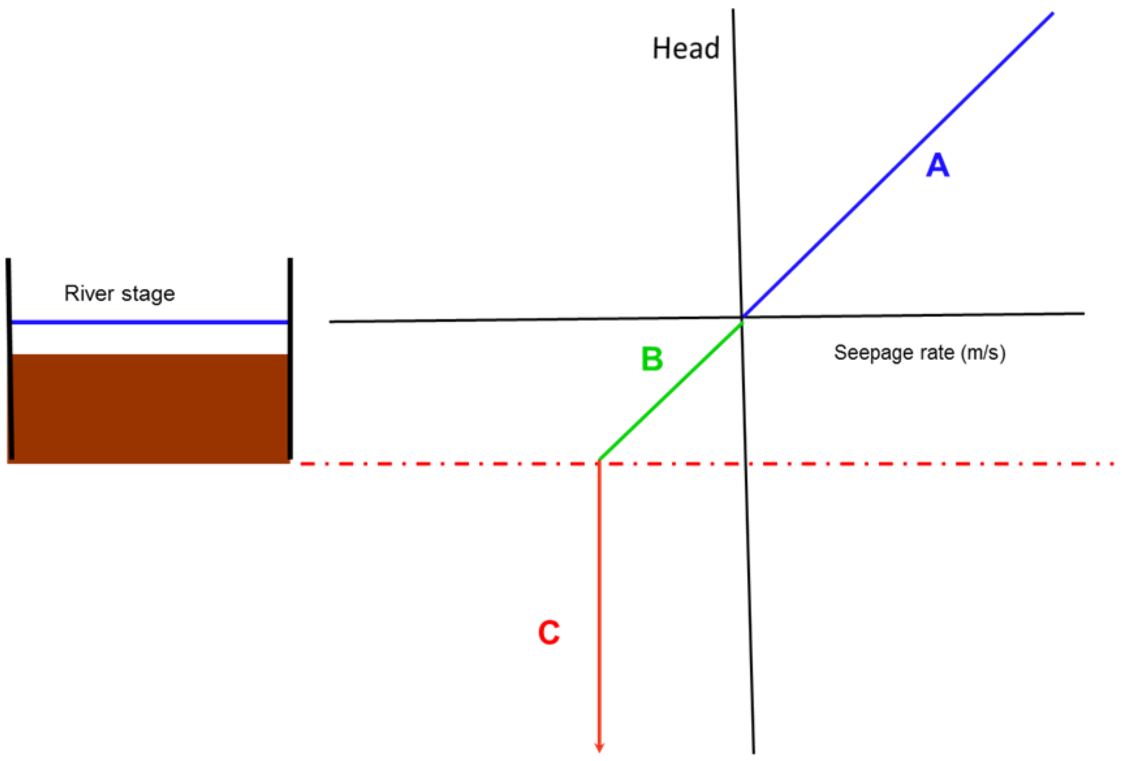


Figure 3-5. Conceptual presentation of stream/aquifer interaction in MODFLOW

Case C in Figure 3-5. represents a disconnected river system, where the downward flux attains a constant value below a critical water table depth (WTcr), i.e. this is what occurs in Zone 1 and Zone 3 (Figure 3-4) in the Namoi valley. In relation to artificial recharge basin schemes, Bouwer (1978) studied disconnected river systems in detail. Due to lack of fast computers in the 1970s, visualisation of flow systems was carried out using an electrical-resistance network analog. Results of this early modelling work suggested that the change from gravity-controlled flow to flow controlled by slope of the water table occurs when depth to water table (DW) is about twice the width W of the recharge system. Furthermore, postulated that the depth of water table has no effect on seepage for condition C in Figure

3-5.. In this conceptualisation at a streambed scale, the soil below the clogged layer will be unsaturated and the flow will be vertically downward at unit hydraulic gradient (Bouwer, 1978). In other words, Bouwer (1978) reached the same conclusion as the more detailed modelling by (Brunner, 2009a).

Hydrological processes leading to groundwater mounding are well studied in irrigation development areas (Petheram, 2008) and in artificial recharge sites These studies are usually linked to irrigation salinity and related land management issues. However, the occurrence of groundwater mounding in ephemeral channels in the Namoi Valley and other Northern MDB catchments are not well documented in the literature. To my knowledge, there is only one case study in literature, where groundwater mound after the 1984 flood in the Lower Namoi was reported (Allen, 2009). Thus, hypothetical simulations were carried out get an insight into factors influencing the development and dissipation of groundwater mounding in connected and disconnected groundwater systems. For instance, the position of water table in relation to streambed elevation has been suggested as a surrogate indicator for classification of groundwater systems into gaining, losing and throughflow systems (Rushton, 2007; Rushton and Tomlinson, 1979).

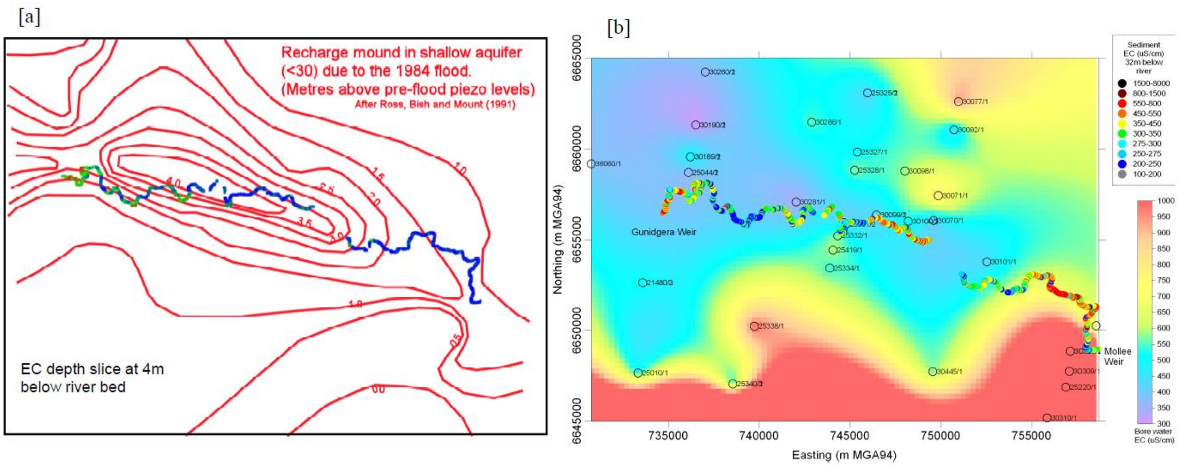


Figure 3-6. Groundwater mound due to the 1984 flood in the Lower Namoi Valley compared with 4 m resistivity depth slice (Allen (2009)

3.2.2 Methods

3.2.2.1 Model Selection and Input Parameters

The USGS VS2D model is the base model for both the VS2DT and VS2DH package. While VS2DT includes solute transport with water, VS2DH concentrates on energy transport. The VS2DT package variably saturated, two- dimensional solute transport model (Healy and Ronan, 1996a; Hsieh et al., 2000b) solves the Richards equation for fluid flow and the advection dispersion equation for solute transport. Results of simulations using the VS2DT package were used to estimate maximum seepage rates for different hydrological conditions. The outcome of these experimental simulations should be considered as a complementary to the field experimental study rather than a standalone methodology. For instance, how does magnitude of percolation varies in depth? Thus, the VS2DT package was a helpful tool to visualise and conceptualise hydrological processes below the streambed. The package is designed based on the unit hydraulic gradient model, where the matric potentials below the active root zone contribute little to the total hydraulic gradient, exhibiting a uniform matric potential profile as illustrated on the hypothetical profile in Figure 3-7.. In this case, $d\Psi/dz = 0$. According to the steady-state unit gradient model, the downward flux below the root zone or fluctuation zone equals the flux across the water table interface, or recharge rate (Stephens, 1996).

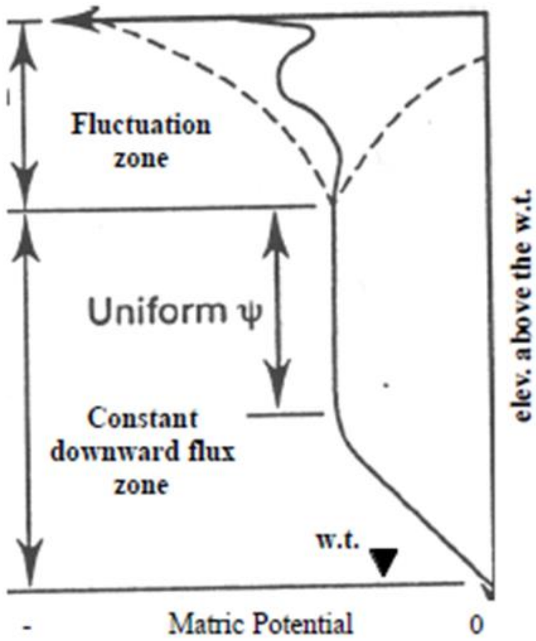


Figure 3-7. Unit gradient model (after, Nimmo et al., 1994)

3.2.3 Model Domain Set-Up: Hypothetical Scenarios

A conceptual model development for hypothetical simulations is similar to previous investigators (Jackson, 2005; Osman and Bruen, 2002; Peterson and Wilson, 1988). A two-dimensional model domain was established which is 30 m deep by 40 m wide (Figure 3-8). For illustrative purposes, a river channel that is 2 m wide and 1 m deep was included in the middle section of the model domain. Model grid cells were set at 0.25 m deep by 0.25 m wide. A ‘steady’/equilibrium state simulation for durations varying between 10 to 50 days was run for different water table depths, until the stream leakage at the bottom of the profile stabilised.

The right and left hand side boundaries were set to a no flow boundary. During ‘steady’ state simulations, the river was set to a zero stage and the lower boundary condition, was set to the depth of the water table (where the pressure head = 0) (Figure 3-8). The model assumptions are listed below:

- The source of infiltration is a rectangular channel with a width of 2 m;
- The soil layer is uniform and isotropic;
- Flow in the unsaturated zone is 2-Dimensional;
- Flow is steady state and governed by Darcy’s Law.

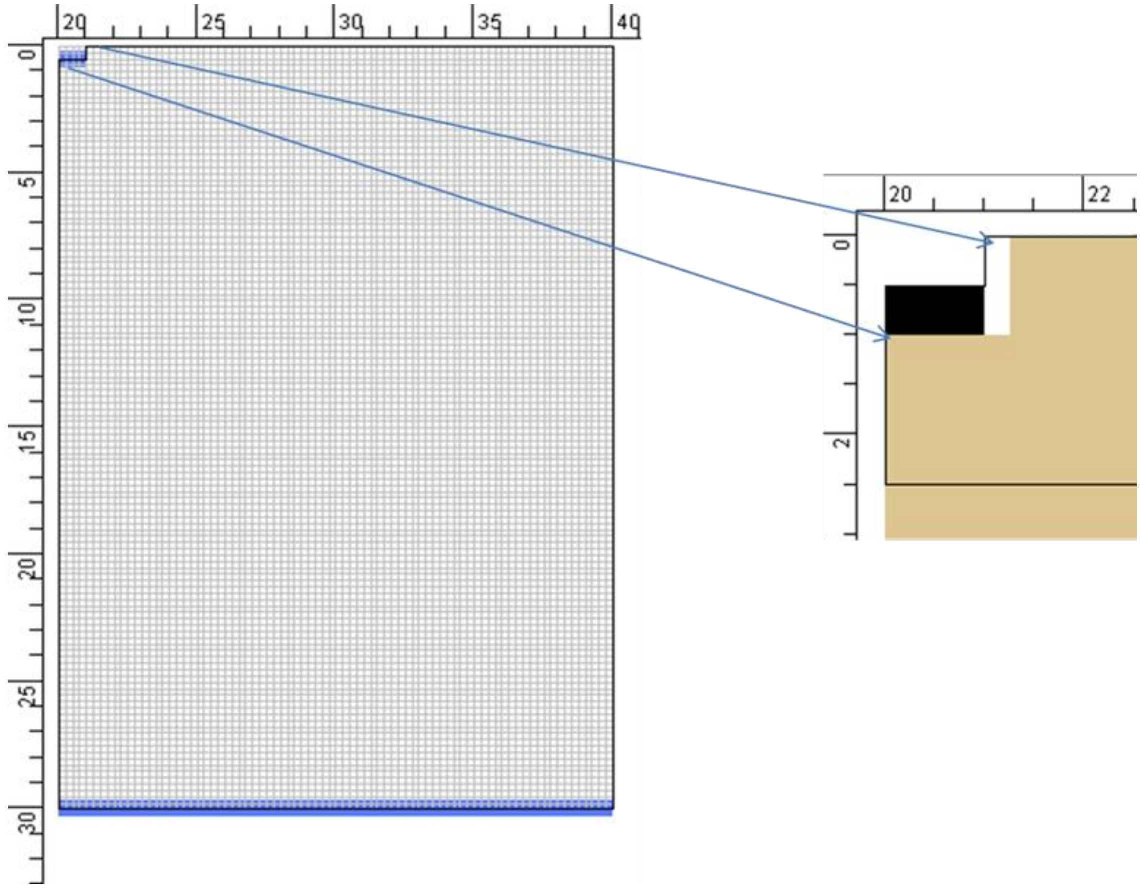


Figure 3-8. A two-dimensional conceptual model used for visualisation groundwater mounding in disconnected river systems. An imbedded sub-figure on the right hand side of the figure, illustrates in detail the configuration of streambed layers (black represents a clogging layer; light brown underlying the clogging layer represents a soil layer with relatively higher hydraulic conductivity). The blue lines represent the upper and lower boundary conditions

The thickness of unsaturated zone or depth to water table provides the volume of empty space available in the unsaturated zone to be filled during the passage of flow or flood events. A thicker unsaturated zone can store more water and can act as a buffer in regulating water from the point of infiltration at the streambed before it joins the groundwater body. When a groundwater mound occurs, a maximum height of groundwater mounds occurs under the mid- point of the channel. The height is very sensitive to the allocated hydraulic conductivity and inversely proportional to the specific yield of the soil material (Karanth, 1987).

Due to lack of real data sets on soil hydraulic characteristics below the stream channel, hypothetical pressure head distribution, for disconnected river systems with water tables of 15 m depth with different combination of hydraulic conductivities of the streambed and the aquifer material are simulated. A 15 m water table depth probably coincides with water levels prior to the commencement of irrigation in the Lower Namoi Valley (Figure 3-8). Based on this hydrological transect, the assertion that the Namoi Valley was a gaining system prior to groundwater development is incorrect and misleading.

The aquifer isotropic hydraulic conductivity was allocated K_x of 1 m/d (fine sands), and corresponding clogging layer hydraulic conductivities were reduced by a factor of 1, 0.1, 0.01 and 0.001 to represent fine textured sediments, varying from silt to clay. Because of symmetry, simulations included only half of the model domain, as shown in Figure 2. Overall an anisotropy ratio of 0.1 was assumed (K_v/K_h = 0.1). Given a fixed streambed configuration, simulations mainly focused on how simulated stream seepage was affected by a stage wise reduction in the clogging layer hydraulic conductivity. In all simulations the stream width was allocated a constant value of 2m. However, this channel width is conservative and varies depending on rainfall events, duration and diversion of stream flow. When the channel carries water, a channel width is expected to occur about 65% of the time in the upper and middle part of the Cockburn Valley. In the end increases in channel width with high flow would only affect the simulations in a minor way.

3.3 Results

3.3.1 Simulated fluxes for Different Scenarios

Simulated pressure head distributions and different combinations of streambed to aquifer hydraulic conductivities are shown in Figure 3-9-to Figure 3-12.The calculated fluxes for different scenarios are presented in Table 3-1. Due to high evapotranspiration and extensive groundwater pumping, the occurrence of groundwater mounding in the Namoi Valley is likely to be a short term temporal phenomenon. Thus, the simulations presented below are considered to be hypothetical.

Scenario I: Both aquifer material and the clogging layer consist of fine sand, where $K_{clog} = K_{aq} = 1$ m/d (1.157×10^{-5} m/s), with anisotropy ratio of 0.1, for both layers (Figure 3-9).

Most of the previous investigators assumed the presence of a clogging layer as a prerequisite for the occurrence of disconnected river system (Bouwer, 2002b; Osman and Bruen, 2002). However, this assumption is not considered valid in the upper catchments of the Namoi River, where the in-stream sediments consist predominantly of boulder, cobbles, and gravels where armoured layers are common in the Upper Cockburn Valley.

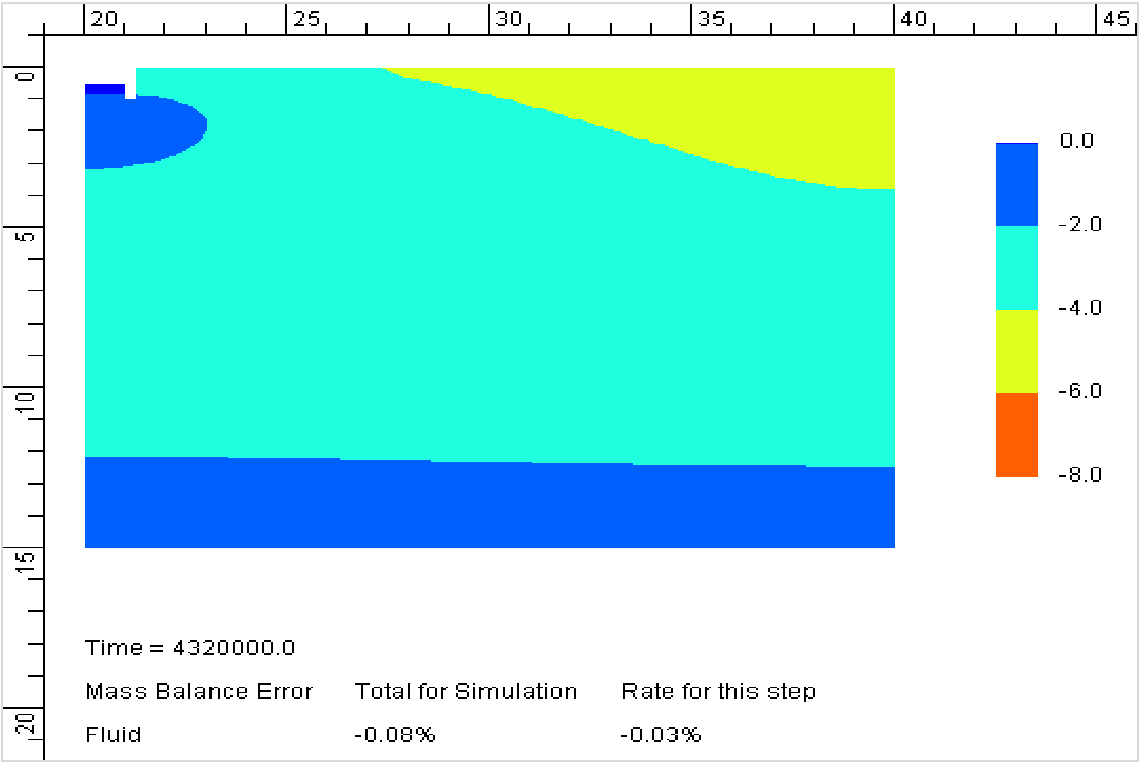


Figure 3-9. Groundwater water head distribution for Kaq= Kclog = 1 m/d.

This scenario also illustrates the general case of homogenous media, where a 'steady' state simulation was carried out for 50 days. The main point is that, even without a clogging layer, a disconnected system occurs as indicated by the negative pressure heads. The pressure head exhibits a bell type distribution, where the influence of the river is evident both in vertical and horizontal directions under the streambed. An oval shaped water table is obvious just under the streambed. However, both its lateral and vertical extents are limited under steady state conditions. In the majority of the model domain the pressure head varied between -2 to -4 m, thus indicating an unsaturated zone. During the simulation, the effects of groundwater pumping and evapotranspiration on the system were not included.

Scenario II: The hydraulic conductivity of the clogging layer is reduced by factor 0.1 or K_{clog} = 0.1 m/d (1.157×10^{-6} m/s), with anisotropy ratio (K_v/K_h) of 0.1, for both layers (Figure 3-10).

In this scenario, the aquifer consists of fine sands with hydraulic conductivity of 1m/d, overlain by a clogging layer with a reduced hydraulic conductivity of 0.1m/d. Clogging layers, which cover streambeds and banks of a river, consist of silt, clay sediments and biologically formed organic deposit.

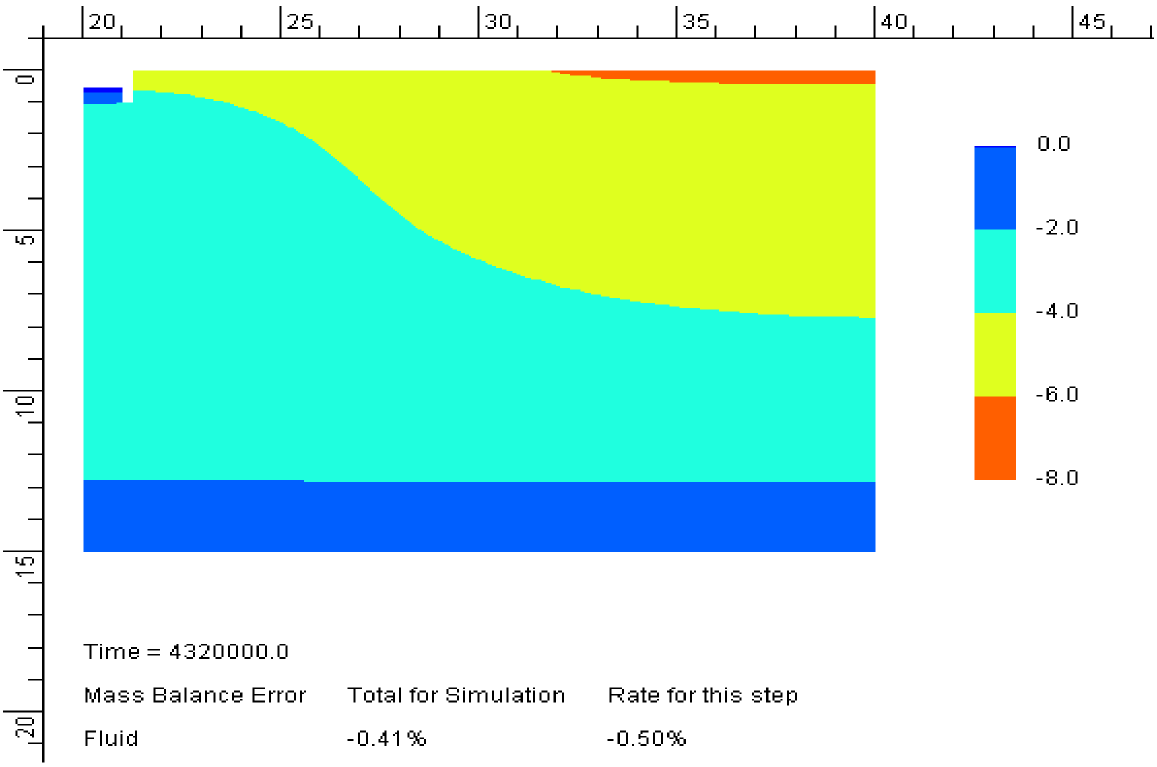


Figure 3-10. Groundwater water head distributions for Kaq= 1 m/d and Kclog = 0.1 m/d

Simulation after 50 days shows a bell shaped pressure head distribution, again with a clearly disconnected groundwater table and an unsaturated zone the majority of the model domain is dominated by negative pressure head between –2 to –4 m below the channel. Even close to the streambed pressure heads up to –2 are simulated. More negative pressure heads are evident on the right hand side of the bank, where the pressure head varies between –8 to –4m.

The propagation of pressure head distribution in the vertical direction is limited when compared with the previous scenario. Groundwater mounding height is insignificant, because the relatively high k value below the clogging layer created favourable condition for lateral movement of water.

Scenario III: The hydraulic conductivity of the clogging layer is reduced by factor 0.01 or K_{clog} = 0.01 m/d (1.157×10^{-7} m/s), with anisotropy ration (K_v/K_h) of 0.1, for both layers (Figure 3-11).

In this particular scenario, as the hydraulic conductivity of the medium decreases, a bell shaped distribution as shown in the above scenario transforms into a linear increment of head with depth, suggesting the dominant flow direction is vertical rather than horizontal; V_z ranges from 4×10^{-6} to 6×10^{-6} m/s. In addition the head distribution in depth doesn't exhibit any resemblance to mounding shape.

The volumetric moisture (water) content after a 100 day simulation is predominantly dry within the model domain, ranging between 0 – 0.2 and relatively high moisture content was observed in a localised area below the streambed, in a narrow zone, above the water table.

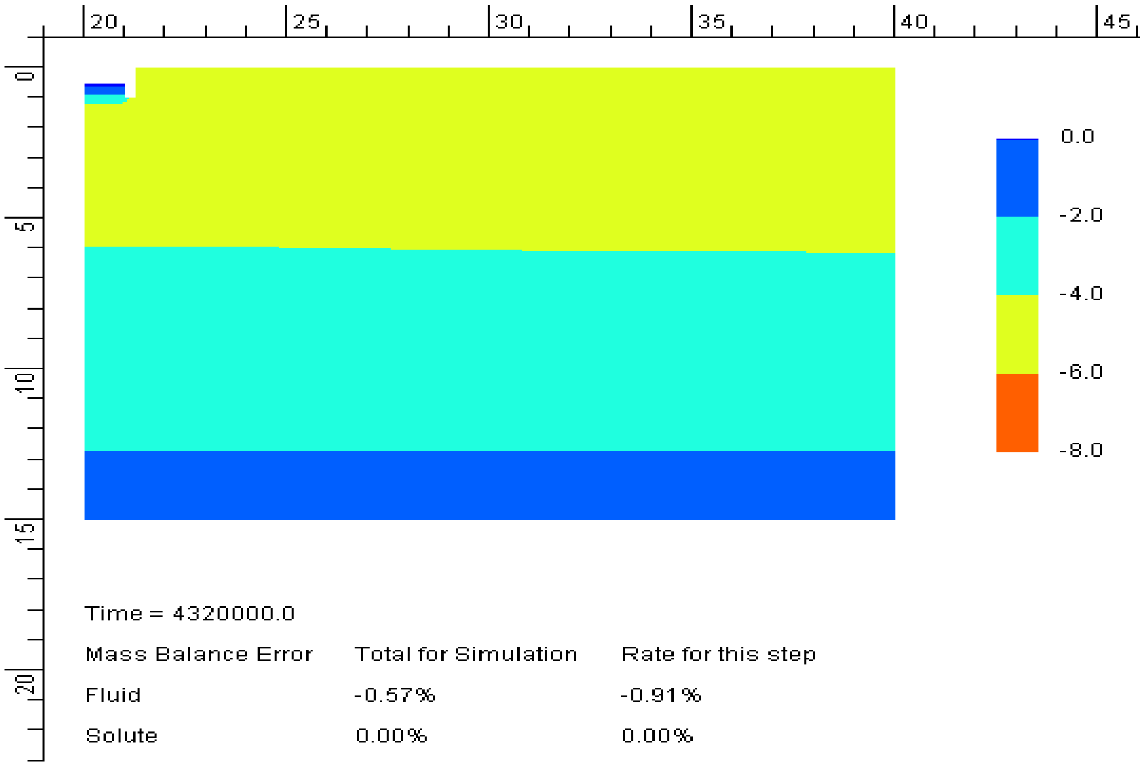


Figure 3-11. Pressure head distributions for Kaq= 1 m/d and Kclog = 0.01 m/d (with anistropy ratio of 0.1)

Scenario IV: The hydraulic conductivity of the clogging layer is reduced by factor of ten or K_{clog} = 0.001 m/d (1.157×10^{-8} m/s), with anisotropy ration of 0.1, for both layers (Figure 3-12).

The vicinity of the streambed shows a higher negative pressure, suggesting a drier domain. Based on the shape of head distribution, the horizontal component pressure propagation diminishes rapidly (Figure 3-12). The horizontal seepage at 14 m in the vertical profile is almost negligible (-2.04×10^{-11} m/s), while the calculated vertical seepage is about 3.97×10^{-7} m/s.

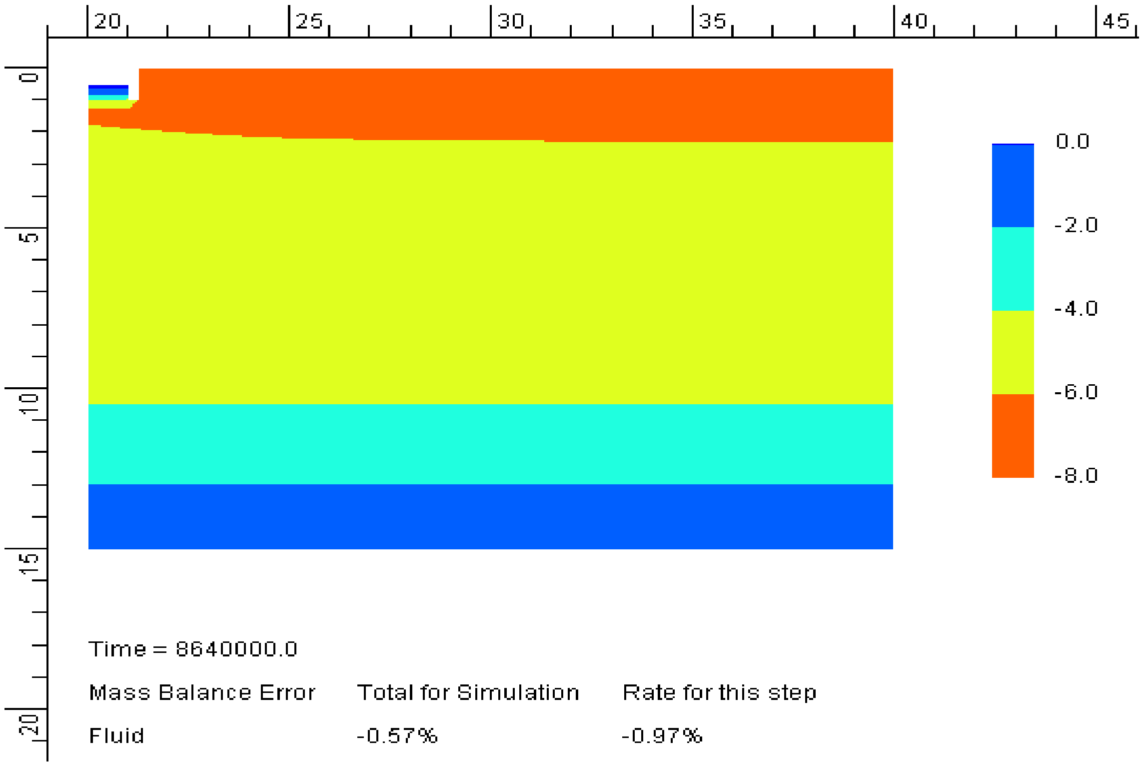


Figure 3-12. Groundwater water head distribution for Kaq =1 m/d and Kclog = 0.001.

This particular scenario may provide a clue on the magnitude of kz, when a channel becomes cutoff or isolated from the underlying saturated zone. This tend to be a characteristic of tight clay soils, which are slow to absorb water and ineffective in storing moisture.

For the hypothetical simulations carried out, vertical seepage ranged from 1.4×10^{-6} to 3.97×10^{-7} m/s for horizontal hydraulic conductivity varying from 1 to 0.001 m/d (Table 3-1) and the relationship between vertical seepage and hydraulic conductivity appears to be linear, with a slope of 0.09. Probably, the derived relationship between V_z and K_x is associative rather than causative relationship.

Table 3-1.Variation of stream seepage with hydraulic conductivity of sediments

Depth (m)	K_x (m/d)	Anisotropy (K_z/K_x)	Horizontal Seepage (m/s)	Vertical Seepage (m/s)
14	1	0.1	1.52×10^{-8}	1.4×10^{-6}
14	0.1	0.1	2.04×10^{-9}	5.0×10^{-7}
14	0.1	1	1.43×10^{-7}	7.2×10^{-6}
14	0.01	0.1	1.04×10^{-10}	4.3×10^{-7}
14	0.001	0.1	-2.04×10^{-11}	3.97×10^{-7}

3.3.2 Determination of the Critical Water Table Depth

The relationship between stream seepage rate and groundwater head was developed using results of steady state simulations (Figure 3-9- to Figure 3-12). In the vertical profile (Figure 3-14, the first point is represented by the hydrostatic condition, where the groundwater head is equal to the stream stage or head. This particular point in the profile represents the upper limit of the streambed, in terms of vertical stratification upwards. Lower in the vertical profile, the 'incipient disconnection' point, defines the initial phase, at which the stream becomes disconnected, i.e. conversion of the system from a connected losing stream to disconnected losing stream. Further along the vertical profile, the maximum seepage incipient point represents the critical water table depth (WTcr), below this point, a maximum downward flux with a unit hydraulic gradient was observed (Benjamin, 1970; Press, 1992). From modelling perspective, at the critical water table depth, the system changes from a variable head boundary condition into a constant flux boundary condition. Some investigators refer to this point as the Maximum Incipient Seepage (Bouwer, 2002a; Fox and Durnford, 2003; Jackson, 2005; Osman and Bruen, 2002).

The first two points in a vertical profile, relating seepage and water table depth are determined by conceptual model setup or field measurements, while the third point in the profile, representing the WTcr, is determined by a series of steady state simulations (Figure 3-9- to Figure 3-12), in contrast to the earlier work by (Brunner et al., 2009). Based on conceptualisation and a series of steady state simulations carried out as part of this visualisation study, two distinct zones can be identified in the vertical profile.

3.3.2.1 A near Streambed Zone (Between Hydrostatic Conditions and Incipient Disconnection)

The near streambed zone is considered to occur in an intermittent stream with water tables within the streambed region. The local water table can drop below the streambed level, as long as the negative head doesn't de-saturate the streambed moisture. In some parts of the Cockburn Valley, this criterion is not always valid. Specifically, during an extended dry spell, the in stream sediments, consisting of boulders and cobbles drain quickly (Figure 3-13). From qualitative field observations, this is particularly true in river gravels and armoured zones of the catchment.

Figure 3-13. The Upper Cockburn is underlain by cobbles, boulders and coarse gravel. In this part of the catchment, the clogging layer is usually represented by an 'armoured zone'.

In this range of water tables, the relationship between water table and stream seepage is non-linear. Previous investigators (Brunner et al., 2009; Fox and Durnford, 2003) sub-divide this zone into two sub- zones, probably based on hydrological conditions in humid and sub-humid environments, where the occurrence of perennial hyporheic flow is a norm rather than exception. However, historical water level data shown in from the Namoi Valley (Figure 3-1-Figure 3-3), is not in agreement with a conceptual model postulated by previous investigators. Based on streambed temperature measurements (Chapter 4), soil moisture contents in disconnected river systems are considered to be low most of the time and don't create a buffer for moderation. Thus, the upper zone above the maximum incipient seepage has been conceptualised as a fluctuation zone (Figure 3-14), which is considered to be highly transient in the inland river systems of the Namoi Valley.

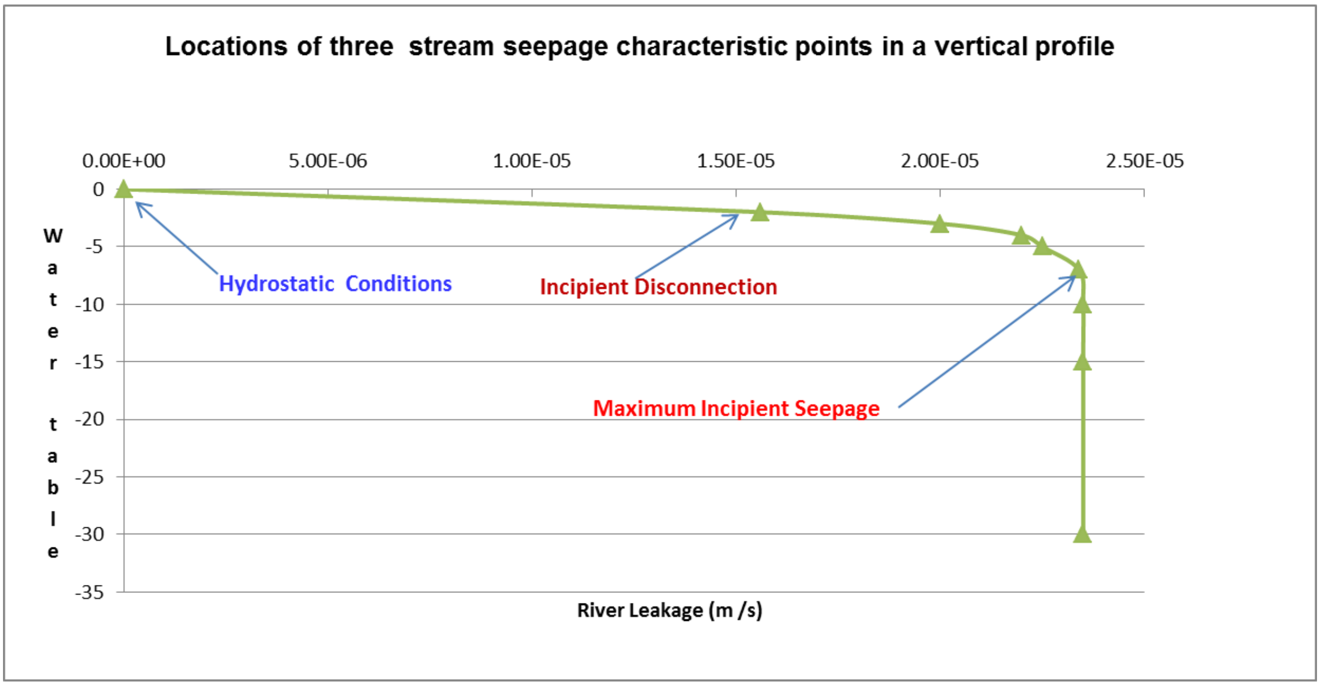


Figure 3-14. Locations of three stream seepage characteristic points in a vertical profile

3.3.2.2 A Far-field Zone (not in the vicinity of the streambed)

A Far-field zone is below the Maximum Incipient Point shown in Figure 3-14. Conceptually, this zone extends from the critical water table depth to the water table, occurring only in an ephemeral river system. Development of groundwater mounding is illustrated using a hypothetical disconnected river system with depth of 15 m below the channel elevation; the distribution of pressure heads and moisture conditions after 100 days equilibrium state simulation is presented in Figure 3-12.

Hypothetical relationships between stream seepage and water tables for different ratios between clogging and hydraulic conductivities are shown in Figure 3-15. As expected, a hypothetical model simulation induced maximum stream leakage, based on a ratio of 10, between hydraulic conductivity of aquifer and clogging layer, when compared to ratios 1 and 5. An absence of a clogging layer, represent a homogenous aquifer, the hydraulic conductivity of the clogging layer is similar to the aquifer hydraulic conductivity with the ratio of 1.

In homogenous aquifers, overlain by a clogged layer, stream seepage increases with increasing the ratio of aquifer hydraulic conductivity to clogging layer hydraulic conductivity (Figure 3-15). However, in areas, where a fine textured material overlies a coarse layer, water moving downwards is impeded under many conditions. In addition, unsaturated flow is complicated by the role of multiple phases (water, air and soil material; Nimmo, 2005).

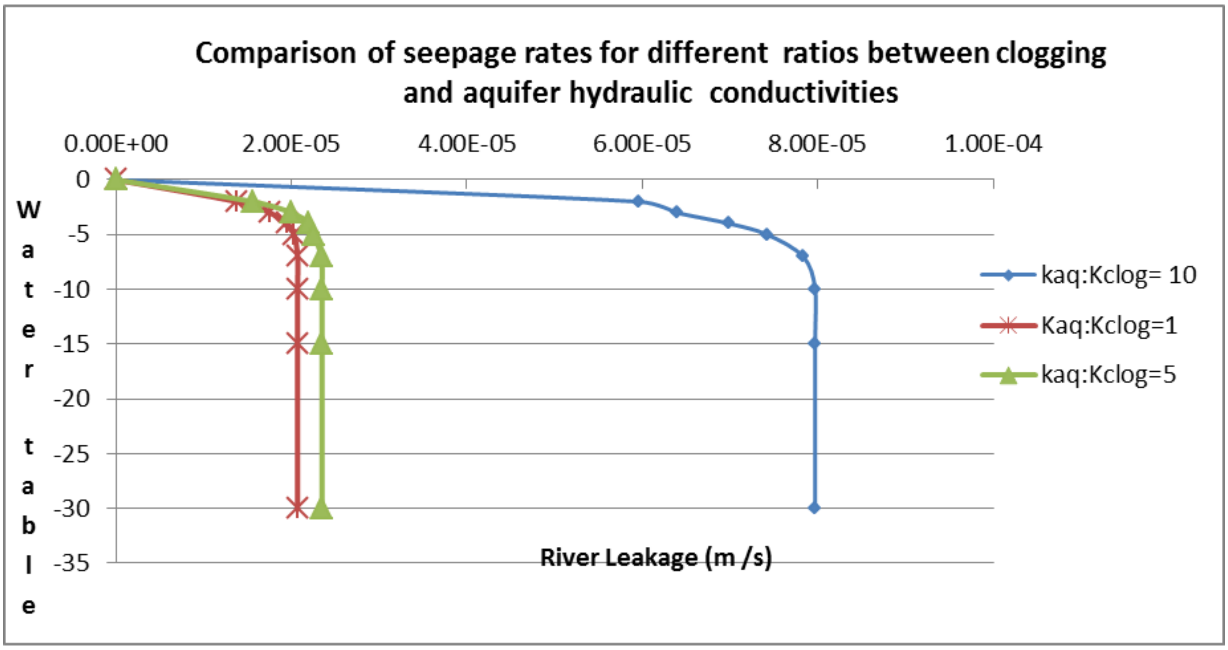


Figure 3-15. Comparison of seepage rates for different ratios between clogging and aquifer hydraulic conductivities

Although there are few field based case studies, which assess the impact of pumping on ephemeral and intermittent streams. Nevertheless, from water resources management perspective, it is important to determine the critical water depth, below which the depth of water table has no effect on seepage rate.

3.4 Discussion

3.4.1 Synopsis

VS2DH was used to estimate hydraulic conductivities and infiltration rates at a streambed scale. The calculated hydraulic conductivities and fluxes for the designated transects are summarised in Table 3-2. Most of the sites are characterised by low hydraulic conductivities and as such the calculated fluxes are considered to be insignificant. For the study sites, the lowest hydraulic conductivity was calculated for the site downstream of the Chaffey dam, while the highest hydraulic conductivity was calculated for the sites in the Cockburn River, where in-stream sediments predominantly consist of cobbles, coarse gravel and sand. In some reaches of the Cockburn Valley, where the in-stream sediments consist of boulders and coarse gravel, the applicability of Darcy law in this type of environment is questionable. Nevertheless, the maximum leakage for disconnected reaches was estimated using a simple Darcy equation of the form:

$$Q_{max} = K_{aq} \cdot W \cdot L \qquad \text{Eq. 3-1}$$

where:

K_{aq} = saturated hydraulic conductivity of the aquifer;

W = width of the river, and

L = length of the river reach.

The maximum calculated river leakages for 2 m width of a river and unit hydraulic gradient are presented in Table 3-2.

Table 3-2. Calculated hydraulic conductivities and fluxes for selected sites

Sites	K_{avg} (m/s)	$Flux_{min}$ (m/s)	$Flux_{max}$ (m/s)	Max Leakage (m^3/s)
D/S of Chaffey	3×10^{-6}	2×10^{-6}	4.3×10^{-6}	6×10^{-6}
Piallamore	5×10^{-6}	1×10^{-8}	1.4×10^{-5}	10×10^{-6}
Baldry	4.5×10^{-6}	3×10^{-6}	6×10^{-6}	9×10^{-6}
Gunnedah	1.22×10^{-6}	1.20×10^{-6}	1.25×10^{-6}	2.4×10^{-6}

3.4.2 The critical water table depth and reversal of fluxes

This study and previous similar hydrological work carried out by (Bouwer, 1969; Brunner et al., 2009a; Osman and Bruen, 2002) suggest that a critical water table depth, where a surface water body becomes disconnected depends on multiple parameters, which include the geometric configuration of the streambed, hydraulic conductivities of the clogging layer and aquifer and depth to groundwater. Based on the parameters used in these experimental simulations, the depth to the critical water table depth (WTcr) was found to be lower than 5 m below channel elevation. This value appears to be appropriate for some transects in the Peel Valley However, each site has its own hydrogeological characteristics and assessment of WTcr should be based on local knowledge and sound hydrogeological judgment rather than a virtual formula.

The above oversimplified presentation of surface/groundwater connectivity used in this did not take into account the reversal of fluxes caused by evapotranspiration from riparian vegetation. Naturally, some of the infiltrated water is expected to be intercepted by groundwater ecosystems. Given the fact that environmental flow is directly linked to surface groundwater connectivity, and is a major water resources and environmental issue in the MDB, the results provide some insight on surface/groundwater connectivity, and the hydrological conditions, where stream depletion takes place.

In the context of Australian environmental conditions, seasonal variation in drainage through the unsaturated zone was studied at the Gnangara Mound, Western Australia by (Sharma, 1991). Average annual drainage for the period of study was 90 mm (13% of precipitation) at the 10 m depth and 70 mm (10% of precipitation) at the 18 m depth. Clearly, the hydrological study carried out at the Gnangara Mound site does not support the constant flux model below the critical water table depth. In addition, the isotropic and homogenous medium (Figure 3-16a) is an oversimplification of a real world, instead a model domain consisting of a layered profile (Figure 3-16b) should be used and a strong potential for lateral rather than vertical flow above lithological discontinuities (Nimmo and Deason, 2002). Field experiments show that volumetric water content and unsaturated flow mechanism in the unsaturated zone vary in a complex manner, the main issue being that the parameter moisture content, matric potential and hydraulic conductivity are interrelated (Nimmo and Deason, 2002). Therefore, a constant flux or piston flow assumption is difficult to defend in the context of this study.

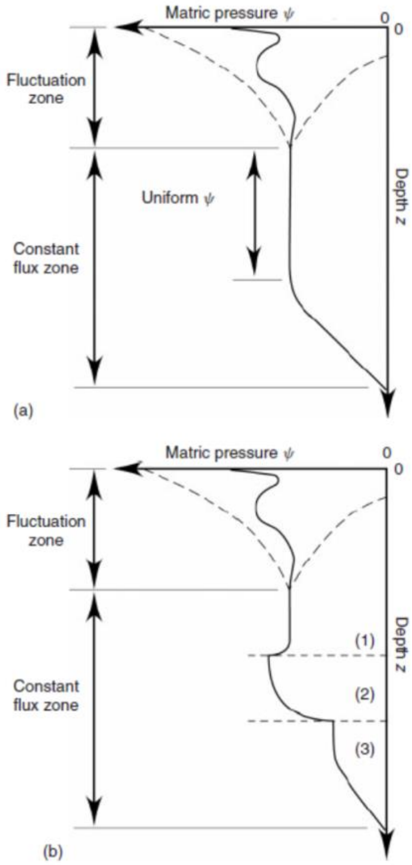


Figure 3-16. Hypothetical profiles of matric pressure as a function of depth in an unsaturated zone deep enough that its lower portion has a constant downward flux of water in (a) a uniform and (b) a layered profile (Nimmo, 2002).

3.4.3 Criterion for disconnection

Due to complexity of hydrological processes of disconnected streams in arid and semi-arid environments, recharge processes of disconnected systems are not well understood. In order to enhance our knowledge of recharge/discharge processes, disconnected streams have been major emphasis of groundwater research activities in Australia. For instance, based on computer simulations, (Brunner et al., 2009a) identified that parameters such as stream depth (*d*), the thickness of the streambed clogging layer (h_c), and the ratio of the hydraulic conductivity of the clogging layer (K_c) to that of the aquifer (K_a) can be used to assess whether or not a channel in arid and semi-arid environments is disconnected. The ratios are expressed in the form:

$$\frac{K_c}{K_a} \leq \frac{h_c}{d + h_c} \qquad \text{Eq. 3-2}$$

The empirical Eq. 3-1 has two major limitations. First, the parameters of natural groundwater regime are controlled by complex interactions between three components of the hydrogeologic environment, namely: (1) climate; (2) topography; and (3) geology (Toth, 2009a). Visual inspection of Eq. 3-1 suggests that disconnection is a function of hydrologic parameters and does not take into account the long term climatic signatures embedded in present groundwater levels. For example, (Jolly and Cook, 2002) identified that for small fluxes lag times between groundwater and surface can be decades to centuries. However, long term climatic signatures are considered to be the main cause of disconnected groundwater systems in arid and semi-arid environments. Similarly, de Vries (2002a) based on a linear reservoir depletion equation showed that the groundwater systems in Botswana, South Africa, required at least thousands of years to manifest a 50 m decline in groundwater levels.

In the context of the MDB, there is a gap of knowledge to assess the impact of climatic changes on past and current groundwater levels. Thus, in the absence of measured climatic records, proxy climatic information can used to construct past climates. For instance, pollen records of past vegetation in eastern Australia suggest that the climate was wetter between 9,000 and 3,500 years ago (Shulmeister and Lees, 1995). Highest water levels are considered to have taken place between 9,000 and 6,000 years ago in Tasmania, between 7,500 and 4,000 years ago in southern mainland Australia and from 5,000 to 3,700 years ago in northern Australia (Shulmeister and Lees, 1995). Probably these changes might have been caused by a regional climatic shift, possibly related to the movement of the subtropical anti-cyclone belt, the westerlies and/or the monsoon (Shulmeister and Lees, 1995). These older climate signals might still be embedded in the current groundwater levels and thus influence the connection between rivers and groundwater.

The second limitation of Eq. 3-1 is related to the dynamic nature of K_c, which has been highlighted in Chapter 5 of this thesis. It is anticipated that a clogging layer breaks up after flood events and K_c approaches K_a, when the ratio on the RHS of Eq. 3-1 transforms to 1, while the ratio on the RHS approaches zero. Thus, Eq. 3-2 becomes invalid from mathematical point of view.

A more detailed insight on the dynamic nature of the clogging layer is provided in Chapter 4 of this thesis. This is an important hydrological issue for stream reaches underlain by aquitards of substantial thicknesses. How does stream leakage takes place via thick clay aquitards, which are characteristics of some catchments in the northern MDB?

3.4.4 Issues of surface/groundwater connectivity in the Peel Valley

From surface and groundwater connectivity point of view the Peel Valley can be divided into two broad regions. The Upper Peel Valley, the area covering upstream of Tamworth, where stream stages are generally higher than groundwater heads. In this area, the reaches are predominantly losing. In contrast, the Lower Peel Valley, which covers an area, downstream of Tamworth, the surface groundwater connectivity issue is more complex for the following two reasons: First, the role of bank storage in augmenting sub-surface discharge increases in the lower Peel Valley, when compared with the upper part of the catchment. Second, the Peel Valley is highly tectonically disturbed and geological features are expected to have influences on groundwater dynamics or preferential groundwater discharge (Figure 3-17). In fact, the author has observed seepage zones from such type of structures, after significant rainfall events.

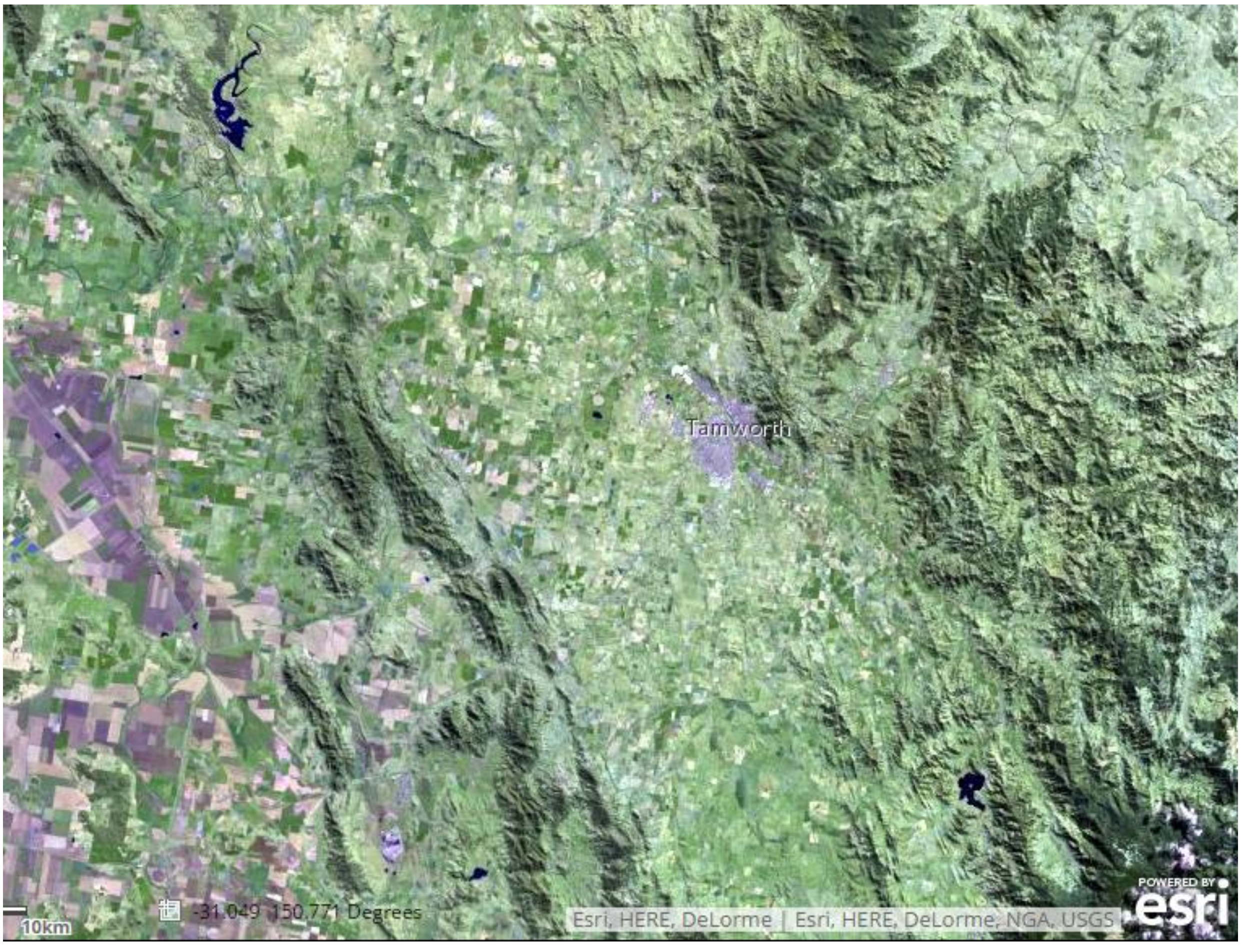


Figure 3-17. Lineaments from a Landsat image in the Upper and the Lower Peel Valley

Chapter 4 Application of Heat as a Hydrological Tracer for Assessment of Infiltration and Near Streambed Water exchanges at two Focus Catchments in NSW

" Among the many problems in heat-conduction analogous to those in groundwater hydraulics are those concerning sources and sinks, sources being analogous to recharging wells and sinks to ordinary discharging wells "

C.V. Theis, 1935

Abstract

Among environmental tracers used in hydrology, heat is considered a robust, inexpensive and effective tracer to assess near streambed water exchanges. This tracer was applied in two contrasting hydro-geomorphological environments in New South Wales (NSW), Peel Valley in the Namoi Catchment, and Baldry in the Central West Catchment. In the Peel Valley, the field investigation extended for more than four years in both regulated and unregulated reaches of the Peel River. The magnitude and direction of fluxes were quantified at the streambed scale. Infiltration is prevalent during low flows, while exfiltration occurs after high runoff events and is determined by small-scale geomorphic features. For instance, in the Cockburn River, representative of an intermittent stream, the estimated infiltration ranged from 0 to 5×10^{-4} m/s. In contrast, at the Baldry catchment the combination of low streambed conductance and infrequent stream flow events of the ephemeral stream has insignificant infiltration rates. Similarly, the infiltration rate downstream of Chaffey Dam in the Peel River is insignificant.

4.1 Introduction

In the selected focus catchments, a comprehensive understanding of the hydraulic linkages between surface water and groundwater is essential for groundwater recharge estimation and improving management of water resources. The streambed zone, or the river-aquifer interface, plays an important role in controlling discharge and recharge and in modifying water chemistry (Desilets et al., 2008; McCallum et al., 2012; Tellam and Lerner, 2009). In the context of the inland rivers systems of the Northern MDB, little attention has been given to the near streambed water exchanges under field conditions. In contrast, most of the hydrological investigations on recharge were carried out in off-channel sites. For instance, in the northern MDB large areas are underlain by heavy clay soils. In these catchments quantification of deep drainage, related to dryland salinity projects, was the main focus of numerous investigations (Silburn et al., 2011; Tolmie et al., 2003). Given the hydro(geo)logical complexity, it is a challenge to estimate recharge to deep aquifer systems overlain by a thick unsaturated zone, consisting predominantly of aquitard layers. For simplicity, in some of the dryland salinity studies, deep drainage was assumed to equal recharge.

In the lowland aquifer systems of the Namoi and the Central West, the regional water table is several tens of meters deep below the channel elevation (Berhane, 2006; Brunner et al., 2011; Lamontagne et al., 2014) Although there are limited case studies, in some riparian zones, infiltrating water may be intercepted by deep-rooted vegetation before it recharges the regional aquifer system (Obakeng, 2007; Walvoord et al., 2002b). Based on the work carried out in different parts of the world, there is increasing evidence that some groundwater dependent ecosystems (GDE) are able to tap water from deep unsaturated zone and groundwater table exceeding 50 m. In addition, evaporation from the groundwater by vapour and capillary transport may take place from more than 20 m depth (Walvoord et al., 2002b; Walvoord et al., 2002a). Evapotranspiration induced upward fluxes may constitute a significant proportion of the groundwater balance (Obakeng, 2007; Phillips, 1963).

Under favourable hydrological conditions, flow pulses in intermittent and ephemeral streams may contribute recharge to the underlying regional aquifers (Constantz, 2008; Fralov, 1967; Izbicki et al., 2008), with the movement of water, energy and momentum are transported from one to another compartment of the hydrological cycle. Thus, following earlier research work (Ronan et al., 1998; Su et al., 2004) changes in temperature related to the energy equation can be used as a surrogate indicator of water exchanges between surface and groundwater systems.

During the last two decades, due to the availability of cheap and reliable temperature loggers, heat as an environmental tracer has gained popularity and is widely used to quantify near streambed fluxes (Constantz and Stonestrom, 2003b; Lapham, 1989b). For example, the occurrence of intermittent and ephemeral streamflow events can be inferred by a qualitative method (Constantz, 2001). This is based on the assumption that the presence of streamflow in an ephemeral channel reduces the amplitude of the diurnal thermal wave propagating through the sediments.

Thus, this work on the application of heat as environmental tracer in selected catchments in NSW was motivated for three main reasons. Firstly, seepage from losing streams or alluvial fans is considered to be the main sources of recharge in the inland rivers of the Northern MDB catchments (Braaten and Gates, 2003a; Lamontagne et al., 2014; Williams et al., 1989). Secondly, the inland river systems of the Northern MDB carry water only in direct response to effective precipitation. The ephemeral/ intermittent nature of the stream systems, in combination with the position of the water table, does not create a favourable environment for horizontal seepage. In fact, there is lack of hydrogeological knowledge on discharge mechanisms at a regional scale. Thirdly, the recent introduction of water trading and release of water to the environment in the MDB requires a better understanding of the components of TL (evapotranspiration, seepages and deep drainage), as this would affect volumes surface water availability. Therefore, the main objectives of this study are to: 1) demonstrate the potential application of the thermal method in ephemeral and intermittent unregulated and unregulated streams in NSW (representing in-channel sites), and 2) estimate the hydraulic conductivity of soils and quantify infiltration rates based on soil temperature measurements at a climatologic station in Gunnedah (055024), representing an off-channel site.

4.2 Study sites

Two focus catchments (Peel Valley in the Namoi Catchment and the Baldry in the Central West Catchment) in NSW were chosen for this study (Berhane et al., 2011). These catchments provided a combination of high stream flow variability and abundance of heat fluxes, which offer ideal hydrothermic environments for the application of heat as an environmental tracer at different spatiotemporal scales and geological complexities.

The Cockburn River, which is unregulated section of the Peel River catchment, originates on the Northern Tablelands of NSW; the Cockburn River is representative of an intermittent stream.

The study site was selected due to its hydro-geomorphologic regime and density of gauging stations. The alluvial deposits in the Cockburn Valley are restricted to a narrow zone between Nemingha and Ballantines Bridge. The maximum thickness can reach up to 15 m. Fractured Carboniferous sediments or the Monibi Granites underly the unconsolidated aquifer system (Berhane et al., 2008).

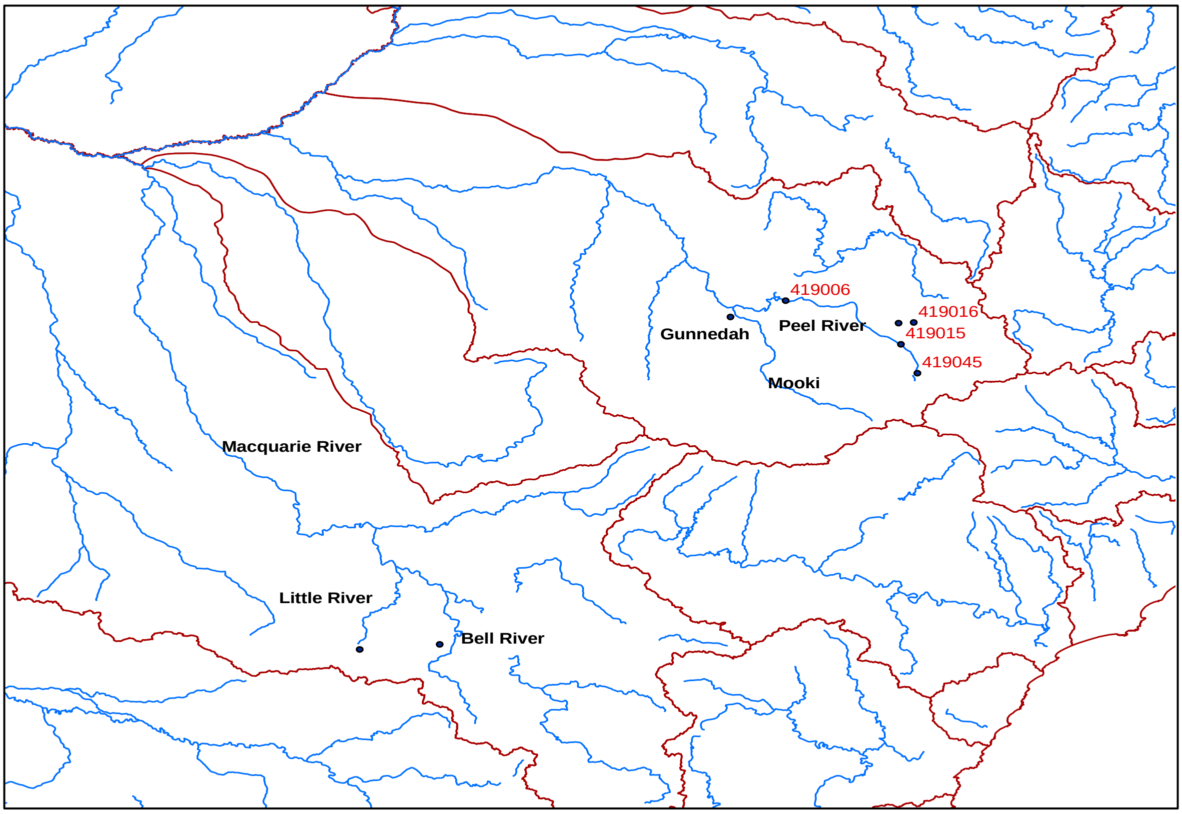


Figure 4-1. Locations of transects/study sites in the Namoi Valley and Central West Catchments (Baldry is near the Little River)

The average annual rainfall decreases with decreasing altitude from the highest point in the Great Dividing Range (1440 m) to Tamworth (280 m) from 800 mm to 650 mm. Tamworth has an annual rainfall of about 670 mm (BOM).

The regulated section of the Peel River is represented by the catchment area, downstream of Chaffey Dam, which is located 35 km to the south east of Tamworth. The catchment lies in a complex geological area known as the New England Fold Belt (NEFB), which represents an assemblage of fault systems (Berhane et al., 2008). To date, the role of the NFEB on the regional groundwater flow system has not been studied. Chaffey Dam commenced storage in

1979, and was constructed to regulate flows in the Peel River for the expansion of irrigation in the Namoi Valley and to augment domestic supplies for the city of Tamworth.

The Baldry Creek catchment is situated south of Dubbo, south west of Wellington and to the east of Goobang National Park. The creek drains into the Little River, a tributary of the Macquarie River. The geology of the Baldry Catchment consists of granites, overlain by regolith, alluvial and colluvial deposits. The regolith, not including creek bed alluvium, can generally be divided into three horizons: the soil zone, saprolite, and a transition zone between saprolite and weathered bedrock. The granites appear to be well fractured and deeply weathered at least in parts (Berhane et al., 2008). The mean annual rainfall at Baldry is 660 mm.

4.2.1 Thermal method and assessment of near streambed water exchanges

Heat is a useful environmental tracer, which has been used for more than 100 years for (Anderson, 2005) estimating infiltration/recharge and discharge rates at different space and time scales. Temperature measurements under streambeds, and in bores located close to the channel, can be used to estimate stream leakage and groundwater discharge (Anderson, 2005; Constantz, 2008; Fralov, 1967; Lapham, 1989b). Heat can be transferred through sediments by a combination of advection and conduction processes. An advective heat transport mechanism takes place with the movement of water and is more prevalent during flood and high runoff events, whereas conductive heat transport is more prevalent in very low or no water-flux conditions. For most hydrologic applications related to infiltration through alluvial sediments, advection is the primary mechanism for the transport of heat by flowing water, and conductive heat transport is regarded as a negligible component of heat transfer.

In addition, periods of flow in ephemeral and intermittent channels can be identified, by detecting temperature oscillations at several depths below a streambed elevation (Blasch et al., 2004; Constantz and Stonestrom, 2003b; Constantz, 2001). At a catchment scale, remote sensing can be useful to estimate river discharge for remote river sites (Brakenridge et al., 1998; Horritt et al., 2001).

During the compilation phase of this thesis, 1DTempPro package (Voytek et al., 2013), a graphical user interface (GUI) was released by the USGS. Pre- and post-processor features of the GUI facilitated to estimate near streambed water exchanges and vertical hydraulic conductivities of the designated transects in the study area.

4.3 Physical parameters

The required model input parameters related to soil physical parameters, volumetric heat capacity of soils, and thermal diffusivity are briefly described below (Table 3-1). Hydraulic conductivity, which quantifies the soil material ability to transfer water, can vary by orders of magnitude from one streambed to another. In contrast, thermal properties vary in a small range of magnitude (Blasch et al., 2007).

4.3.1 Volumetric heat capacity of soils

The volumetric heat capacity C of soil is defined as the change in heat content of a unit bulk volume of soil per unit change in temperature in Joules per cubic meter per degree. As such, C depends on the composition of the soil’s solid phase (the mineral and organic constituents present), bulk density, and the soil’s wetness (Table 4-1). The value of C can be calculated by addition of the heat capacities of the various constituents, weighted according to their volume fractions as given by (de Vries, 1963).

$$c_b = x_w c_w + x_o c_o + x_m c_m + x_a c_a \qquad \text{Eq. 4-1}$$

Where, *w*, *o*, *m* and a represent water, organic solids, mineral solids and air respectively. Thermal conductivity *k*, is the measure of a material's ability to conduct heat. It is defined as the amount of heat transmitted per unit time per unit area per unit temperature gradient. Units of thermal conductivity are watts (joules per second) per square meter per degree Celsius per meter (W m^{-1} $°C^{-1}$).

4.3.2 Thermal diffusivity

Thermal diffusivity is the ratio of thermal conductivity to volumetric heat capacity. The units of thermal diffusivity are metres squared per second (m^2s^{-1}). Thermal diffusivity is a measure of how quickly an imposed change in temperature is transmitted through the material. Air has a large thermal diffusivity, despite having a low thermal conductivity, because its volumetric heat capacity is small. With almost no capacity for storing or releasing heat, temperature signals travel quickly through air (Constantz and Stonestrom, 2003a).

Thermal properties were assigned to the soil/geologic units as listed in Table 4-2. The thermal properties were assumed to be independent from the hydraulic properties. (Stonestrom and Constantz, 2003).

Table 4-1.Thermal conductivities and diffusivities for selected fluid, solid and mixed phases (Sources:(de Vries, 1963)

Phase	Density (kg/dm3)	Thermal Conductivity (W/m °C)	Specific heat Capacity (kJ/kg°C)	Thermal Diffusivity (10-6 m^2/s)
Water at 5°C	1	0.5724	4.202	0.13622
Water at 10°C	0.9998	0.5820	4.192	0.13886
Water at 20°C	0.9982	0.5984	4.182	0.14335
Water at 30°C	0.9957	0.6154	4.178	0.14793
Sand (dry)	1.2-1.6	0.600	0.800	0.62500
Fine Sand (dry)	1.635	0.627	0.760	0.50459
Fine Sand (saturated)	1.635	0.627	0.760	0.50459
Fine Sand (saturated)	2.02	2.750	1.419	0.95940
Gravel (dry)	1.745	0.557	0.766	0.41671
Gravel (saturated)	2.08	3.070	1.319	1.11900

Table 4-2. Parameters of the numerical model (VS2DH)

Parameters	Value
Saturated hydraulic conductivity (K) at 200C	model output
Porosity	0.3
Volumetric heat capacity of dry solids (ρsCs)	1.2×10^{6} $Jm^{-3}K^{-1}$
Thermal conductivity of the solid-fluid system at fully saturation (Kfs(θfs))	1.6 $Js^{-1}m^{-1}K^{-1}$
Volumetric heat capacity of fluid (ρfCf)	4.2×10^{6} $Jm^{-3}k^{-1}$

In the context of this project, the VS2DH package was used estimate hydraulic conductivities and near streambed fluxes for the designated transects, where three categories of field data sets were used:

- Hourly stream temperature and stage data as an input to the upper boundary condition;
- Hourly sediment temperature measured at 40 and 80 cm depths below channel level;
- Available hourly groundwater level (GWL) and temperature data monitored in off channel piezometers.

Based on streambed thermographs at 40 and 80 cm depths, it was only possible to estimate infiltration rates for designated transects of the study area, but not percolation and recharge rates (Chapter 5 of this thesis). In order to 'track' or visualise the status of infiltrated water below the streambed, the complementary VS2DT package was used next to the VS2DH package. A further explanation of this commonly used concept in unsaturated zone hydrology is provided in Section 3.2.3.

4.3.3 Seepage and Hydraulic Conductivity Estimation

The USGS VS2DH (variably saturated, two- dimensional heat transport model) was used to calculate near-surface streambed exchanges from temperatures measured in the stream and below channel elevations. A detailed discussion of the theoretical foundation for VS2DH and its use can be found in the works by (Healy, 2010; Healy and Ronan, 1996b; Hsieh et al., 2000b). VS2DH simultaneously solves the variably saturated flow and energy transport equations:

$$\frac{\partial[\theta C_W + (1-\Phi)C_S]T}{\partial t} = \Delta \cdot K_t(\theta)\nabla T + \nabla \cdot \theta C_W D_h \nabla T - \nabla\theta C_W \nu T + QC_W T \qquad \text{Eq. 4-2}$$

Where θ[-] is volumetric moisture content; K_t is thermal conductivity of the water and solid matrix; D_h is the hydrodynamic dispersion tensor; ν is water velocity; Q is rate of fluid source; and T is temperature of fluid source.

The left hand side of the equation represents the change in energy stored in a volume over time. The first term on the right hand side describes the energy transport by heat conduction through the bulk material. The second term on the right hand side accounts for thermomechnical dispersion. The third term on the right side represents advective heat transport, and the final term on the right hand side represents heat sources and sinks to mass movement into or out of the volume. The thermomechanical dispersion tensor is defined as (Healy and Ronan, 1996b):

$$D_h = \alpha_T |v| \delta_{ij} + \frac{(\alpha_1 - \alpha_t) v_t v_j}{v} \qquad \text{Eq. 4-3}$$

Where α_l and α_t are longitudinal and transverse dispersivities, respectively (m); δ_{ij} is the Kronecker delta function; v_i and v_j are i^{th} and j^{th} component of the velocity vector, respectively (m s-1), and $|v|$ is the magnitude of the velocity vector (m s-1).

Three-dimensional, variably saturated isothermal fluid flow in a heterogeneous medium is described by an equation that is derived by incorporating Darcy's law into the conservation of mass equation for fluid flow (Lappala et al., 1987). Darcy's law for variably saturated media can be expressed as:

$$v_i = k_r(\psi) \frac{k_i \rho g \partial h}{n_e \mu \partial x_i} = \frac{K_i(\psi) \partial h}{n_e \partial x_i} \qquad \text{Eq. 4-4}$$

Where,

i = coordinate indices (i=1,2,3);

v = average pore velocity of the fluid [L/T];

$k_r(\psi)$ = relative permeability of medium (soil type) as function of pressure head, ψ (dimensionless);

ψ = pore-water pressure head (ψ=h-z) [L];

Z = elevation head [L];

K_i = intrinsic permeability of the medium [L2];

ρ = density of fluid (water) [M/L3];

g = gravitational acceleration [L/T2];

h = total hydraulic head (equal to the sum of the average pore-water pressure head, ψ, and the elevation head, z [L];

n_e = porosity of the medium (void volume/total bulk volume) [dimensionless];

μ = dynamic viscosity of water [M/LT];

x_i = spatial coordinate [L]; and

$K_i(\psi)$ = hydraulic conductivity as a function of pore water-pressure head [L/T].

4.3.4 Analysing model error

A better fit between measured and simulated thermographs was achieved both qualitatively and quantitatively. There are several basic statistical measures of goodness of fit (Benjamin and Cornell, 1970). In this chapter, we have used the root mean square (RMS), which is one of the widely used diagnostic statistics. The root mean square (RMS) is calculated as follows:

$$RMS = \left(\frac{1}{n} \sum_{i=1}^{n} (O_i - S_i)^2 \right)^{0.5} \qquad \text{Eq. 4-5}$$

Denote O_i and S_i as the i^{th} terms of the observed and simulated time series comprising n data points.

4.3.5 Results of Case Studies Based on Field Experimentation

4.3.5.1 Case Study I: Peel Valley (Downstream of the Chaffey Dam)

Between 2000 and 2012, water and sediment temperatures were measured on opportunistic basis at the Chaffey Dam, and stream temperature trends can be visualized from earlier reports (Preece, 2004; Thoms et al., 1999). Water temperatures in the Peel River directly above the dam fluctuate annually between about 7 and 28 °C. Water quality studies showed that water temperatures in the Upper Peel River, along a reach extending about 6 km upstream and downstream of the dam site, were generally more uniform in autumn and winter than in spring and summer. This reach of the river also appeared progressively cooler in the downstream direction during the warmer months. The reversal in stream temperature trends could be caused by cooling caused by vegetation shade (Bek, 1986).

The intake of the Chaffey dam is located within the thermocline (Preece, 2004). Under normal operating conditions discharge is typically small. A combination of small discharge and shallow withdrawal depth minimises cold water pollution. Because of the reservoir height, cold water pollution is probably not an issue at Chaffey Dam (Preece, 2004).

In regulated catchments, heat-tracing method can be used for detection of seepage from an embankment dam (Johansson, 1997; Merkler et al., 1989). Measurements are normally performed manually in open observation wells once a month in order to follow the seasonal variations (Johansson et al., 2000a). Heat is transported by conduction and advection. In the absence of leakages, downstream of a dam, temperature changes are determined by conduction only. However, localised leakages induce temperature anomalies caused by advection.

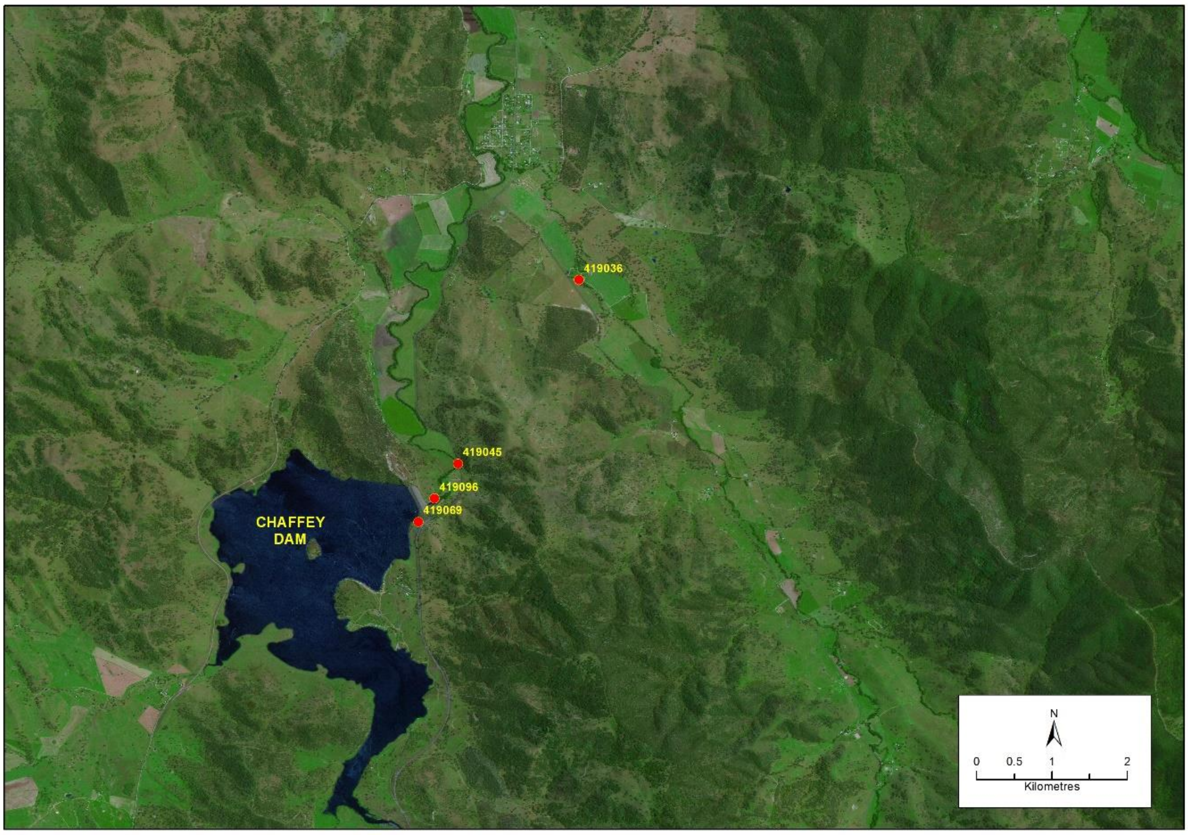


Figure 4-2. Gauging station 419045 is located downstream of the Chaffey Dam

Although the available temperature and EC data is limited at the Chaffey Dam transect (Figure 4-2), an attempt was made to estimate infiltration rate downstream of the Chaffey Dam, in close proximity to the gauging station 419045, where stream temperature and stage is measured on 15 minute interval. Streambed temperatures were measured at 40 and 80cm below streambed elevation. During the observation period, stream temperatures varied from 8.8 in winter to 31.5 °C, while stream stage varied from 0.95 to 4.29 m. As water is released from different levels of the dam, the stream temperature signal was considered to be a composite of temperatures from different depths. Total water release from the Chaffey Dam, during the study period ranged from 2.28 to 8551 MLD, while stream stage and stream temperatures varied between 0.99 to 1.95 m and between 11.35 to 30.96 °C respectively (Figure 4-3).

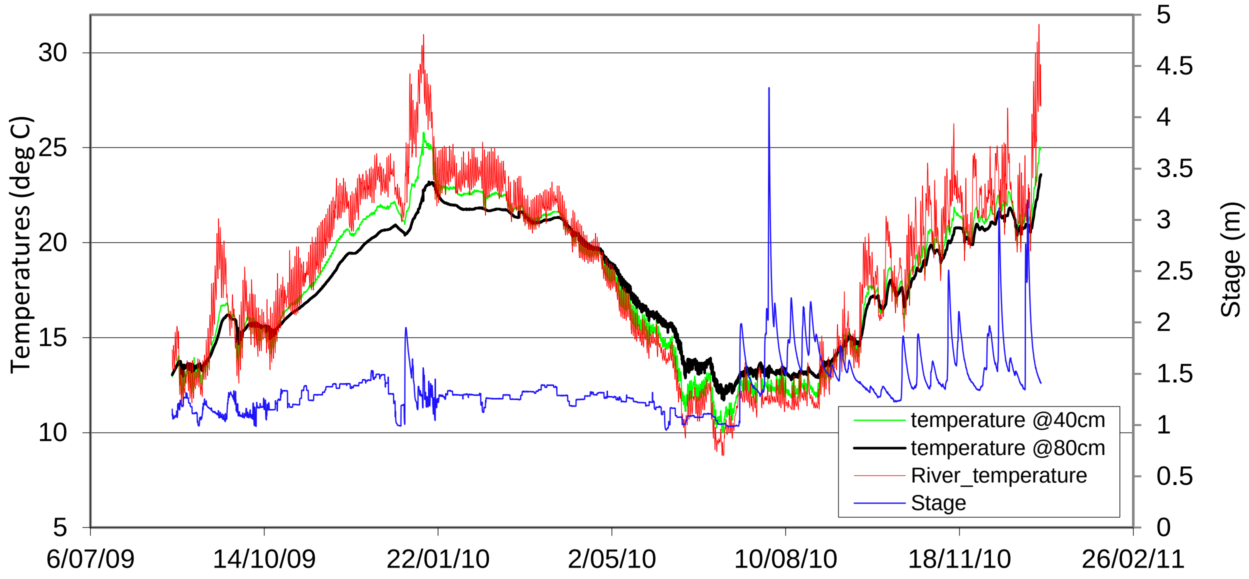


Figure 4-3. Observed stream stage and streambed temperature trends, D/S of the Chaffey Dam

A one-dimensional model was constructed for the period between 25/08/09 and 13/03/10. The model domain was kept as simple as possible in accordance with the available of hydrometric data. The upper part of the model domain, which directly interacts with the Peel River was modelled as a specified, variable head boundary. It was assigned measured hourly stream temperatures and stages at gauging station 419045. The LHS, RHS and the bottom of the model domain were represented as no-flow boundary.

During this the simulation diurnal temperature fluctuations were muted and streambed temperature increased by almost 10 °C (Figure 4-4). Keeping all thermal properties constant, the best fit between measured and simulated streambed temperature at 80 cm depth was obtained for hydraulic conductivity value of 3×10^{-6} m/s; calculated downward flux for the same period ranged from 2×10^{-6} to 4.3×10^{-6} m/s (Figure 4-4).

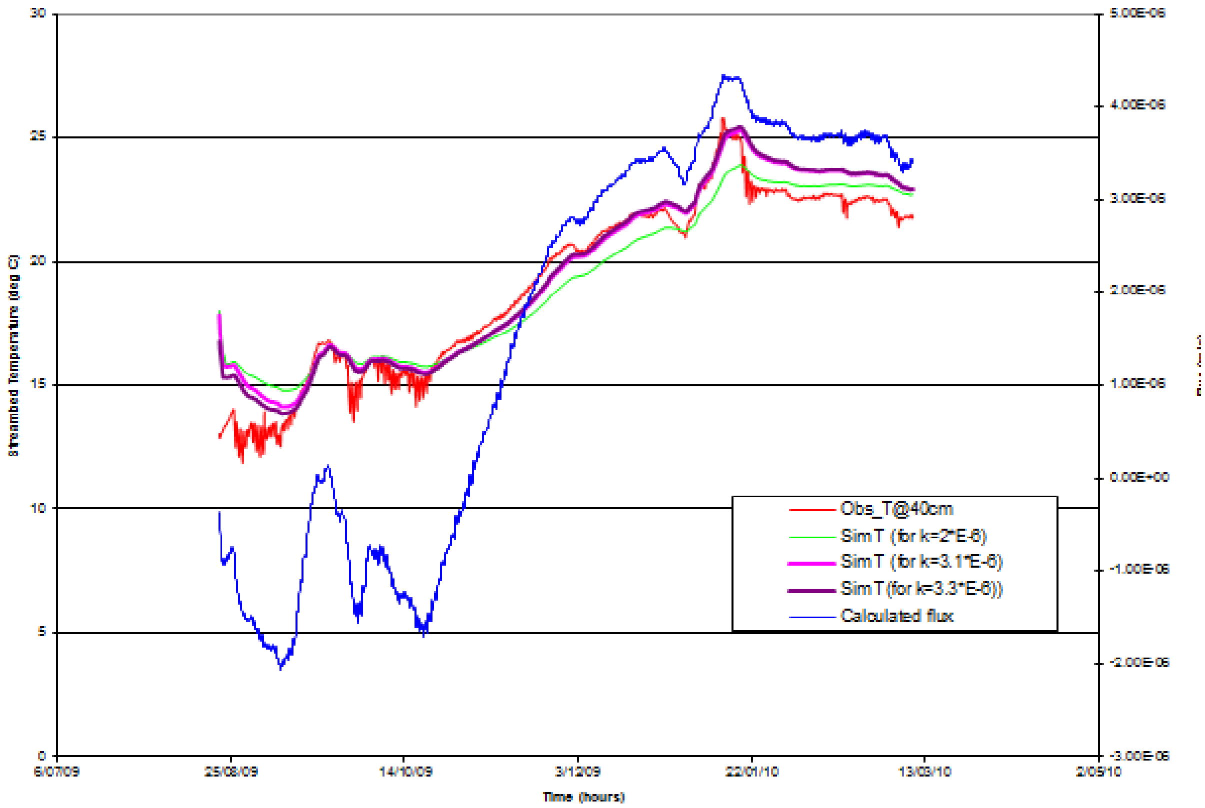


Figure 4-4. Simulated VS observed streambed temperatures at 40 cm depth: D/S Chaffey Dam

4.3.5.2 Case Study II: Peel Valley (Piallamore transect)

4.3.5.2.1 Hydrographs

The Piallamore transect is located between Chaffey Dam and Tamworth. This particular site was an experimental site to study surface groundwater connectivity in early 1990s by the NSW Office of Water (Figure 4-5). Thus, some of the unpublished hydrologic data sets collected in 1990s were used to complement the thermal data set collected during a field campaign of 2008-2012.

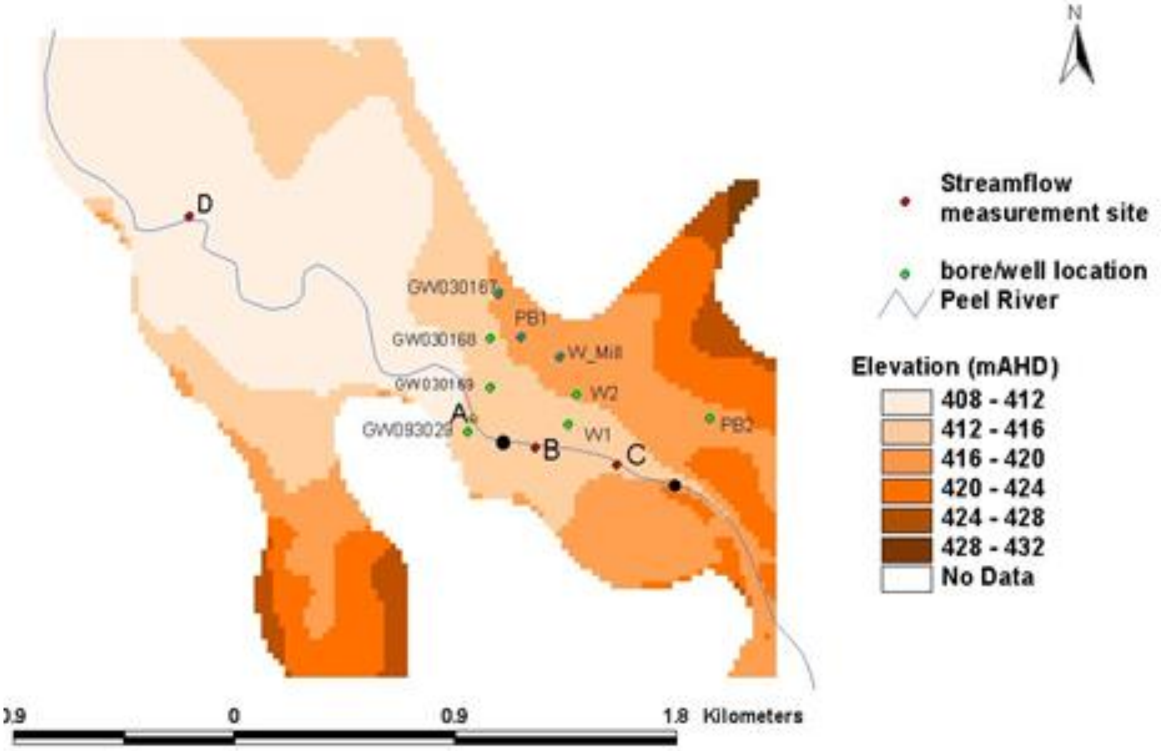


Figure 4-5. Location map for the Piallamore transect

Hydrographs for monitoring bores 30168-2 and 930029, located at the Piallamore transect are shown in Figure 4-6. In all of the monitoring bores, groundwater levels were lower than streambed elevation, suggesting a losing hydrologic regime. However, during flood events of 2004, the groundwater levels were close or above the streambed elevation, suggesting a switching hydrologic regime may occur.

Based on an integrated monitoring of stream stage and groundwater levels, groundwater responses for the 2004 flood events were captured in a transect of monitoring bores along the Piallamore transect. At the Piallamore Gauging Station (419015), the river stage reached heights of 1.65 and 2.37 metres above the pre-rainfall river stage on the 17th and 26th January 2004 respectively. Increased groundwater levels were recorded at all piezometers (Figure 4-5). Hydraulic diffusivity was calculated based on groundwater level responses in close proximity to the Piallamore gauging station. Piezometers GW30168 and GW93029 are 273 and 37 metres respectively from the banks of the Peel River. River stage reached maximum on 17th January at 10:15 am. A lag time 48.8 hours was measured between the time the river reached maximum and the groundwater level in GW30168-peaked (Figure 4-6). The calculated diffusivity at GW30168-2 was 43 m2/d. A storativity value of 0.1 yields a transmissivity at 430 m2/d. This value was estimated using one flood event, therefore, it should be considered as an indicative rather than a face value.

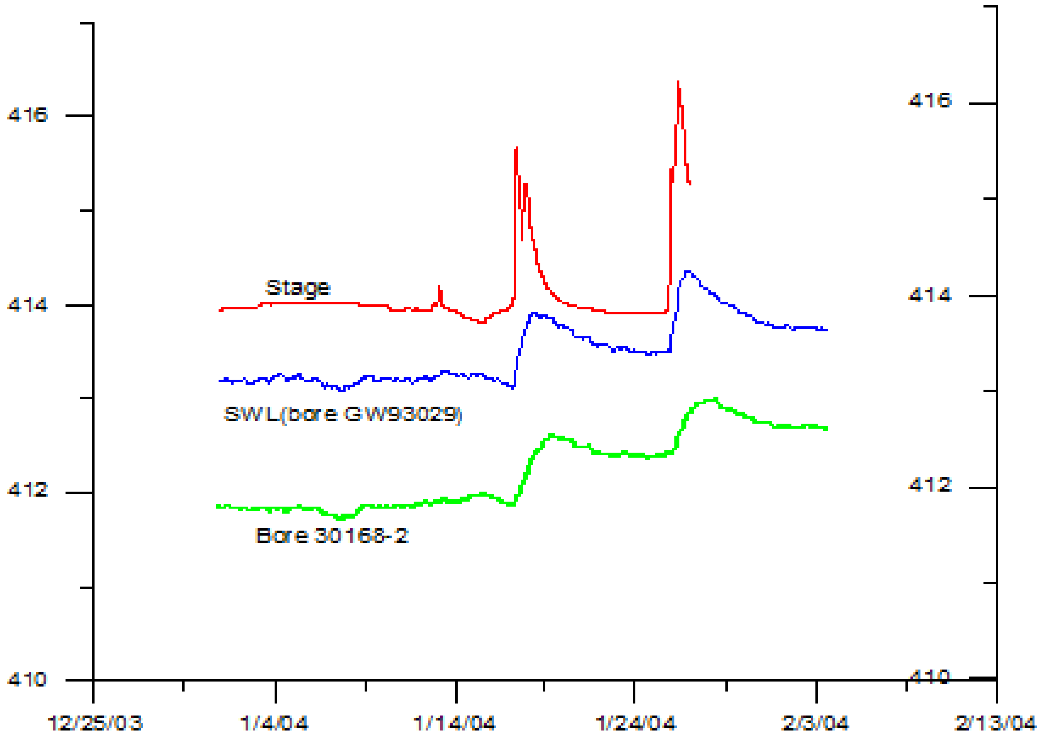


Figure 4-6. Groundwater level trends for monitoring bores at the Piallamore transect

Table 4-3: Variations of river stage and ground water level during the flood events of 2004

Date	Time	Stage (mAHD)	Bore	SWL (mAHD)	t_{lag}
17/01/04	10:15	415.649			
18/01/04	9:00		GW93029	413.842	22.75
19/01	11:00		GW30618-2	412.551	48.75
26/01/04	8:45	416.369			
26/01/04	22:00		GW93029	414.261	12.25
28/01/04	9:00		GW30168-2	412.971	48.15

In contrast, groundwater temperature responses in GW3068-2 were muted, during the two consecutive flood events of 2004 (Figure 4-6), suggesting that the heat wave in comparison of pressure wave is retarded faster (Figure 4-7). The impulse like temperature response observed in 30168-2 is probably caused by a sudden flux of warmer surface water into the bore casing. Nevertheless, the groundwater temperature increased by about 0.1°C in the period between 17/01/04 and 3/02/04, as a result of both distributed and focused groundwater recharge.

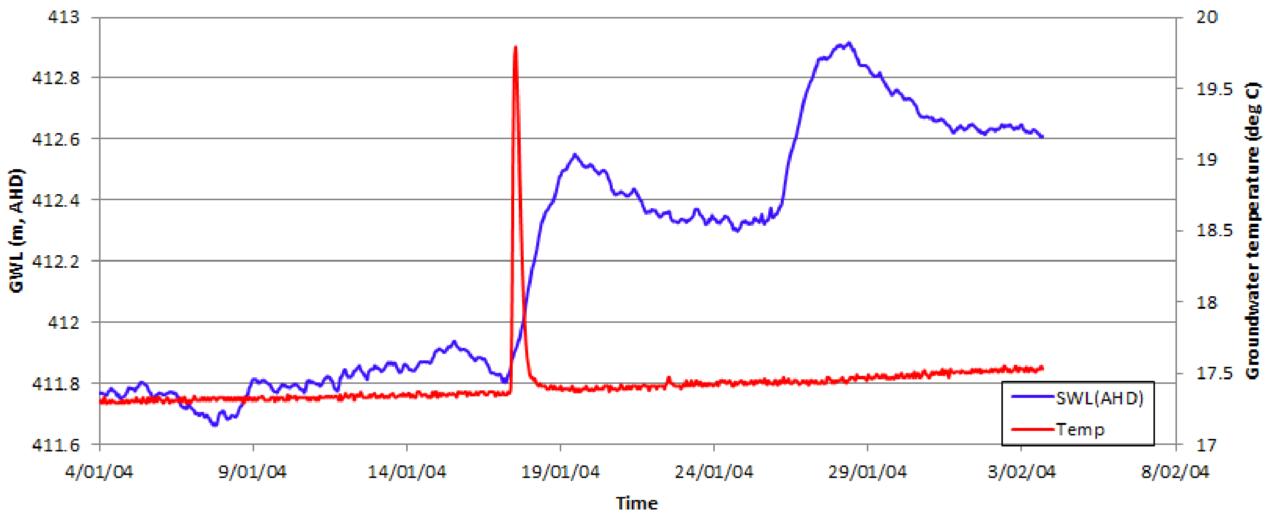


Figure 4-7. Groundwater temperature responses in 30168-2 during the flood events of 2004.

4.3.5.2.2 Thermograph trends

During the monitoring period, stream temperatures varied from 8.8 in winter to 31.5 °C in summer, while stream stage varied from 0.95 to 4.29 m. Both stream and streambed temperatures were lower in summer of 2012 when compared with 2010-11. The decrease in streambed temperatures in the late part of the monitoring period may be linked to changes in burial depths of the sensors caused by sedimentation. However, the streambed temperatures also decreased in sympathy with stream temperatures in the same period (Figure 4-8).

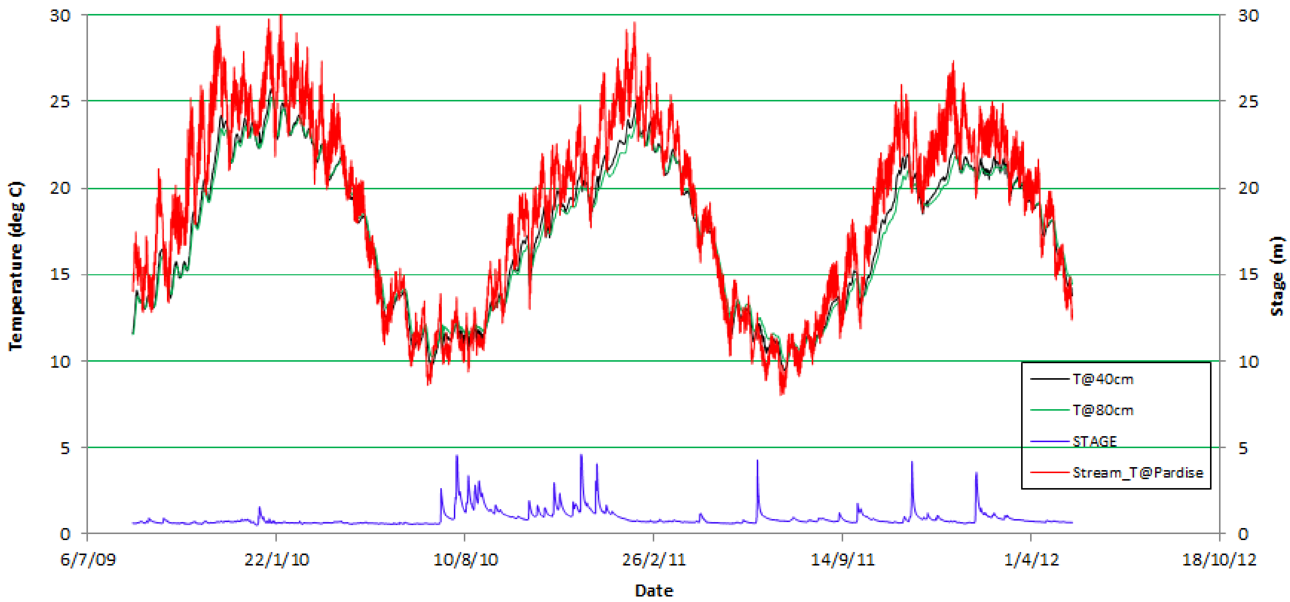


Figure 4-8. Observed stream stage and streambed temperature trends, at the Piallamore transect

Keeping all thermal properties constant (Table 4-2), the best fit between measured and simulated streambed temperature at 80 cm depth was obtained for hydraulic conductivity value of 6×10^{-6} m/s; calculated downward flux for the same period ranged from 3×10^{-6} to 8×10^{-6} m/s (Figure 4-9). Due to depositional environment downstream of the Piallamore transect, there is high uncertainty in the calculated hydraulic conductivity and flux values.

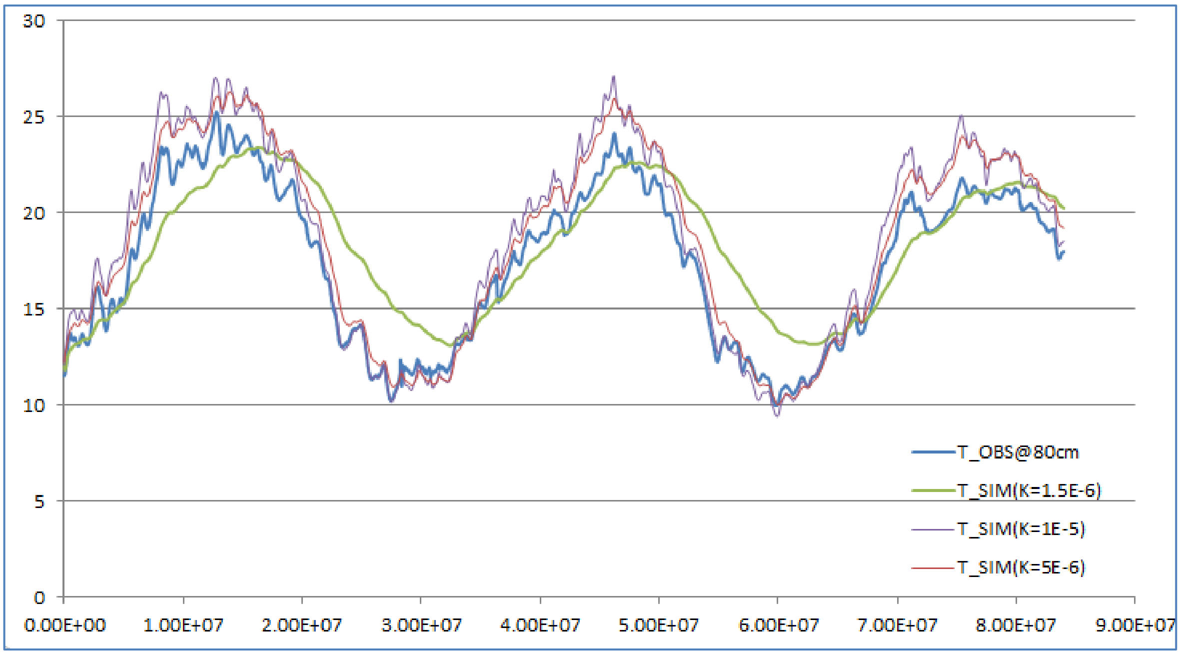


Figure 4-9. Simulated VS observed streambed temperatures at 80 cm depth: Piallamore transect

4.3.5.3 Case Study III: Cockburn Valley (Ballantines transect)

At the Ballantines transect, the maximum width the alluvial sediments extends up to 2 km. The in-channel environment consists of well-developed riffle-pool sequences (Thoms et al., 1999). However, immediately downstream of Ballantines Bridge the river channel is very shallow and the gravel sediment has been completely removed, exposing bedrock material. There are signs of active channel erosion at this site that are associated with sand and gravel extraction (Thoms et al., 1999). Due to the gravel armouring layer, the instrumentation at this site survived the two consecutive flood events of 2008.

In the armoured zone near the Ballantines, we have used VS2DH to estimate the VHC and fluxes. During the two-consecutive flood events of 2008-9, the VHC varied from 8×10^{-5} to 2×10^{-3} m/s, with stream leakage ranged from 9×10^{-5} to 2×10^{-4} m/s, which are 1-2 orders of magnitude higher than in the Peel River.

4.3.5.4 Case Study IV: Cockburn Valley (Smith transect)

This transect is located downstream of the Ballantines Bridge and its midway between the Ballantines and the Kootingal transects, where detailed instrumentation was set up. This transect represents a mobile zone geomorphology in the Cockburn Valley (Chapter 5 of this thesis).

This site is similar to the Ballantines transect; however, channel instability was more evident. Many in-channel features were eroded (Figure 4-10). A privately-owned bore is equipped with a thermistor. Unfortunately, the thermograph contains noise caused by the regular operation of a stock and domestic pump.

Figure 4-10. Location of an in-stream piezometer in the vicinity of the Smith's transect.

In the vicinity of this site the river channel is very shallow and the gravel sediment has been completely removed, exposing bedrock material. There are signs of active channel erosion at this site that are associated with sand and gravel extraction (Thoms et al., 1999). The in-stream substratum consists predominantly of gravel, cobble and sand. Due to the presence of gravel armour layer, the streambed sediments are less prone to erosion when compared with the banks of the river, during, during low flow events.

An in-stream piezometer equipped with thermistors and EC logger was washed away during the flood events of 2008. Thus, groundwater temperature collected in a privately production bore, was used for estimation of hydraulic conductivity and fluxes. For this exercise, a 2D model was constructed for the Smith transect, which consists of 63 and 42 cells. Groundwater temperatures used for simulations are from 25/11/2008 to 31/12/2008. The upper boundary condition was allocated stream temperature and stage data from the Kootingal transect.

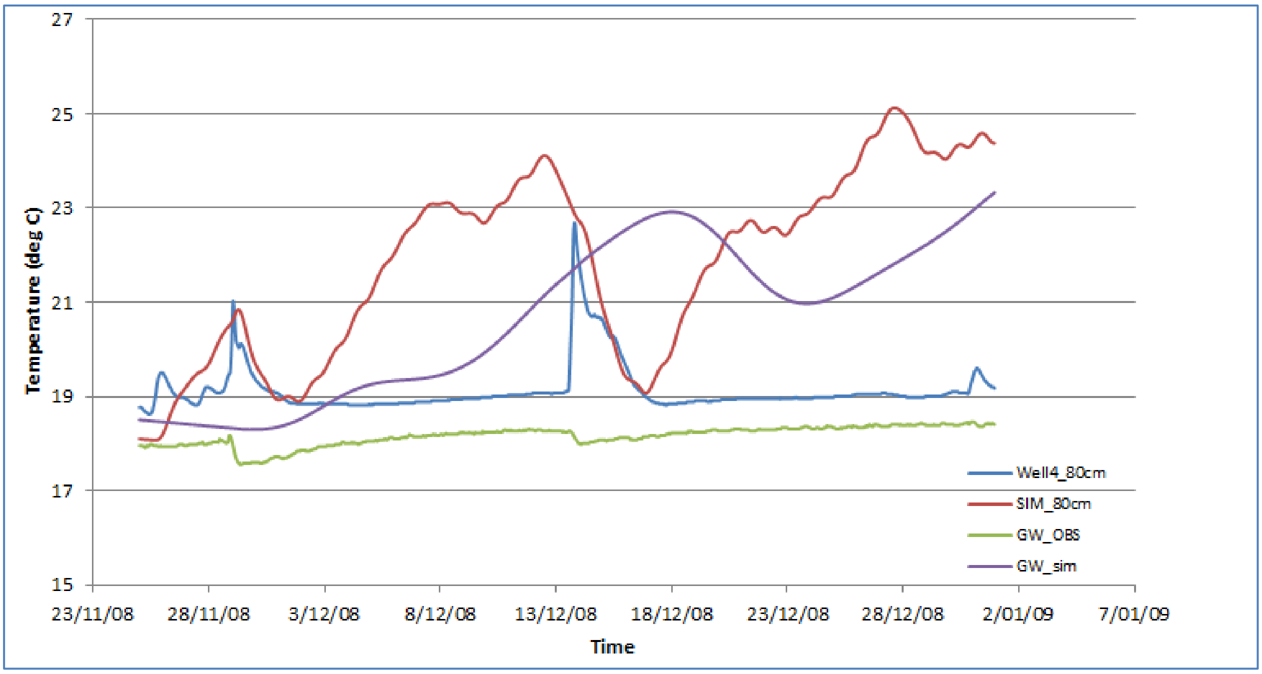


Figure 4-11. Observed VS simulated streambed temperatures at 80 cm depth and groundwater temperature: Smith's transect

shows that the comparison between observed and simulated groundwater and streambed temperatures measured in the vicinity of production Well 4. Remarkably, at this site the streambed temperature exhibits a positive anomaly response, while the groundwater temperature showed a negative anomaly response for both consecutive flood events (Figure 4-11). The difference between simulated and observed temperatures is less during runoff events. However, the discrepancy increases during the recession period or the river ceases to flow. In addition, the bore is used for a domestic water supply and this may have contributed to the discrepancy between measured and simulated groundwater temperatures. In addition, the thermal properties may not be constant during low groundwater levels induced by pumping.

Nevertheless, during the simulation period infiltration varied from 5.94×10^{-6} to 1.96×10^{-5} m/s; although the best fit between observed and simulated was carried out with K_z of 2.6×10^{-6} m/s, due to mobility of the upper zone, hydraulic conductivity should be considered as a highly variable parameter (Chapter 5). A detailed analysis on variability of streambed conductivity is not warranted, due to lack of measured stream temperatures on site.

4.3.5.5 Case Study V: Cockburn Valley (Chapmans Well transect)

A detailed analysis for thermal data collected at the Kootingal transect (419099) is provided in Chapter 5 of this thesis. However, in this section, analysis of thermal data collected in one of the abandoned wells (Well 1) at the Kootingal Well field area is provided. As Well 1 is located in the vicinity of two production-wells, propagation of heat and water are highly likely to be induced by groundwater pumping.

For the period, which includes two consecutive flood events (22/10/2008 to 10/02/2008), groundwater levels varied from 9.53 to 8.47 bgl, while groundwater temperatures varied from *18.9 to 19.4 °C (Figure 4-12). Hydraulic and groundwater temperature responses were 0.3m and 0.2 °C respectively.

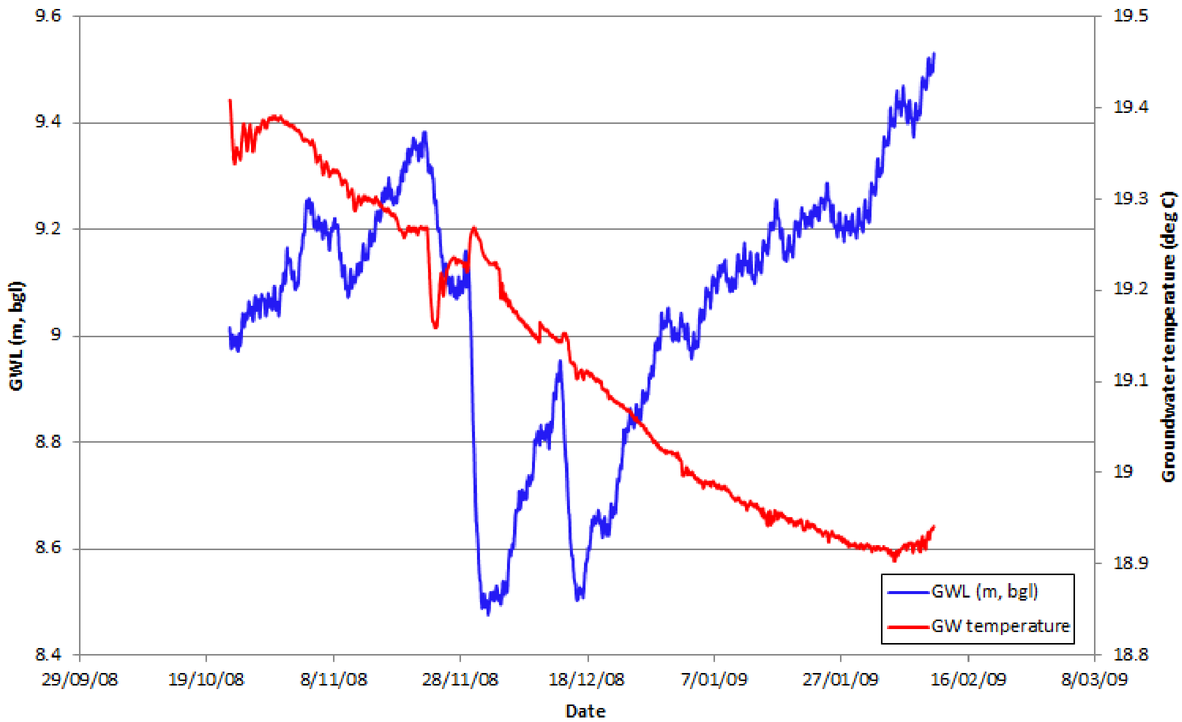


Figure 4-12. Flood induced groundwater temperature response in Well-1

Although this particular well not ideally constructed to measure water level responses during flood events, GWL data collected at this site, provided an excellent example on the degree of connectivity between surface water and groundwater systems in the Cockburn Valley and. Figure 4-13 illustrates a quick response of the groundwater system to runoff and flood events.

In hydrology, continuous wavelet transform (CWT) could be useful in different types of applications:-in river regime characterisation, CWT is used to detect how discharge is related to climatic variability indices (Labat et al., 2004; Nakken, 1999) or to qualitatively analyse how certain characteristics of meteorological input time series are transferred to hydrological output (Gaucherel, 2002).

Thus, in the context of this study, non-stationary relationship between surface and groundwater heads was explored using wavelet analysis presented in Figure 4-14, which is better equipped to study time series data at different frequencies (for non-stationary harmonic analysis). Temperature could be used to infer percolation rates through time series analysis Earlier analysis based on Fourier analysis is provided in an earlier report (Berhane et al., 2008).

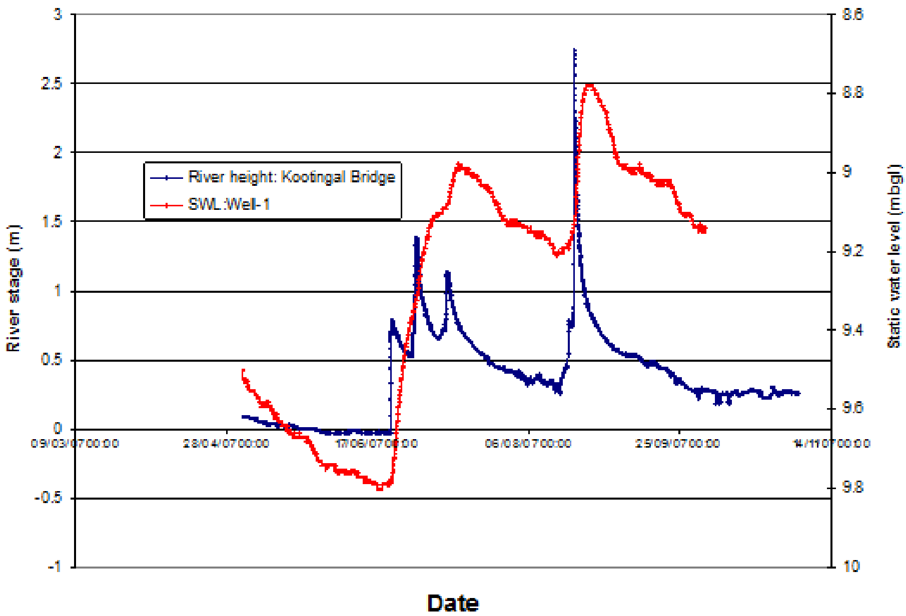


Figure 4-13. A quick groundwater response to runoff and flood events: Cockburn River

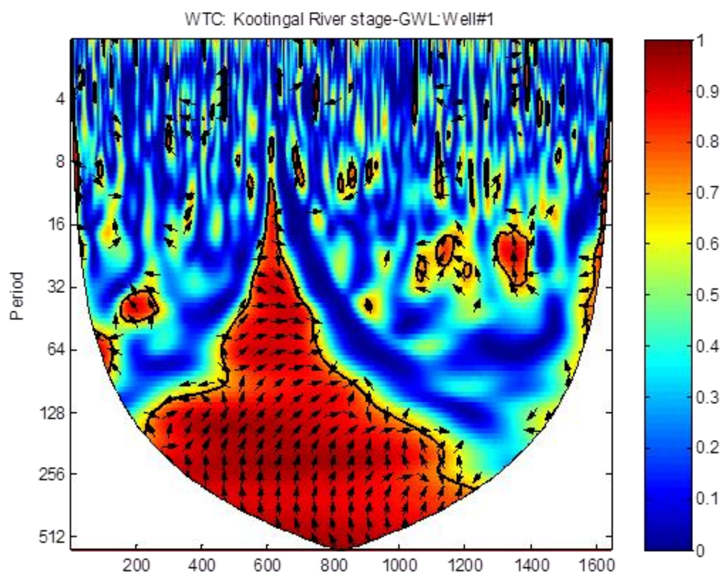


Figure 4-14. Squared wavelet coherence between surface and groundwater heads. The phase relationship is indicated by the direction of the arrow (in-phase pointing right, anti-phase pointing left, and runoff leading GWL by 180º, pointing upward direction).

The best fit between measured and simulated was achieved using Kz value of 6×10^{-5} m/s and simulated vertical seepage ranged from 0 to 1.93×10^{-5} m/s. The RMS values at 0, 0.8 and 9m depths were 0.149, 0.782 and 0.001 respectively (Figure 4-15). Due to the dynamic nature of streambed conductivity, the fit between measured and simulated stream bed temperatures at 80 cm depth were poor. More elaborate simulations, which take into account the dynamic nature of streambed conductivity is carried out in the next chapter of this thesis.

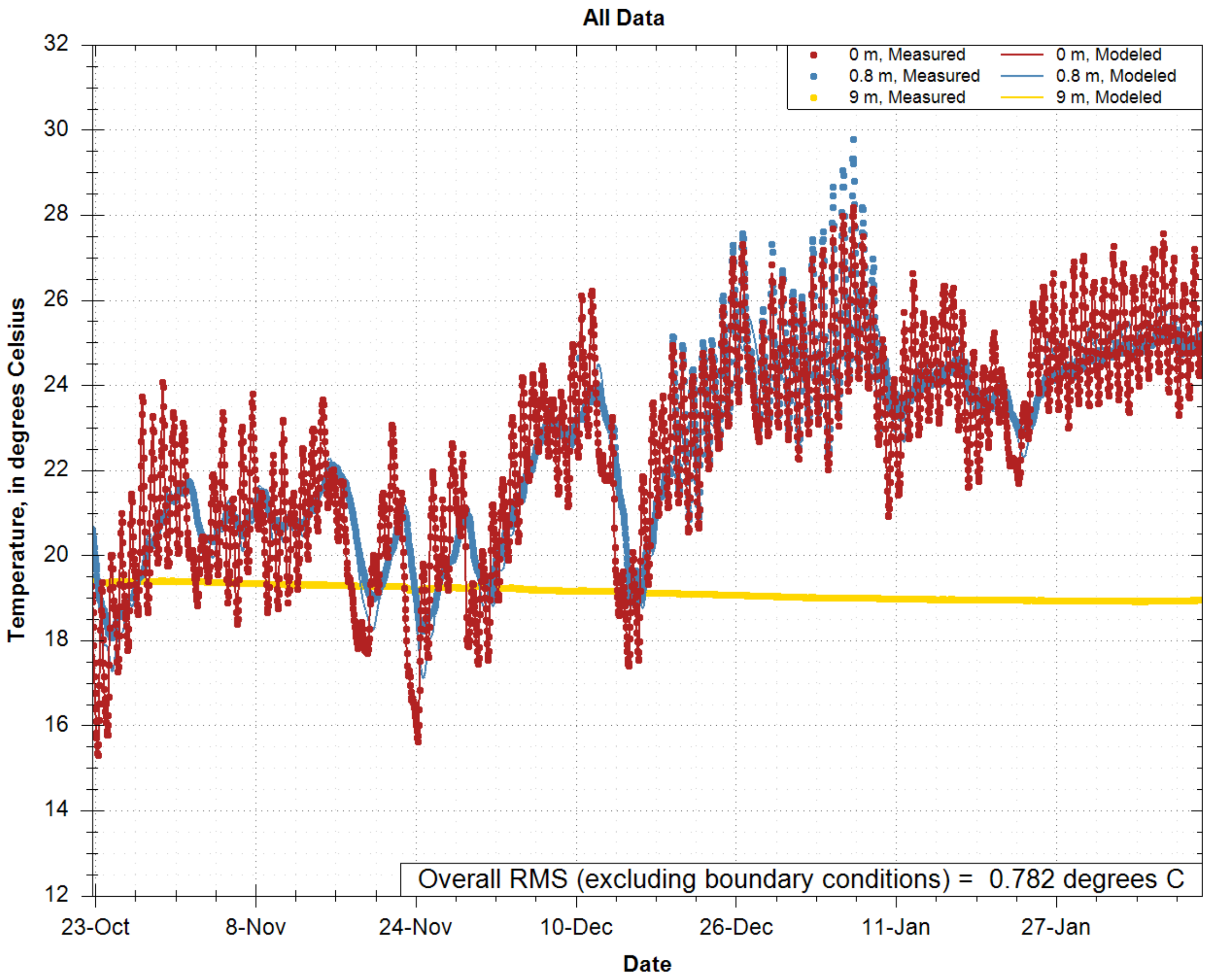


Figure 4-15. Measured Vs Simulated streambed temperature trends at Well-1

4.3.5.6 Study VI: Peel Valley (Carroll Gap transect)

The surface groundwater connectivity in the vicinity of the Carroll Gap is different when compared to the reaches in the Upper Peel Valley., Figure 4-16 illustrates the hydrogeomorphological setup of this part of the Peel Valley; this area is considered to be favourable for the occurrence of riverbank discharge. Near Somerton Village, on both sides of the road cutting, an inversion of topography is observed, where ‘old’ gravel beds cover westerly dipping mudstone of the Devonian and Early Carboniferous age. It is reported that about 30 million years ago, an immense amount of unconsolidated sediments was removed from the Lower Peel flood plain and this may have caused an inversion in topography, where older coarse gravels are located at higher altitudes than the current Peel River flood plain in the vicinity of Somerton (Bob Brown, pers, Communication, 2009: ozgeotours.110mb.com). From the author's field observation in the Peel Valley, this area appears to be favourable for river bank discharge.

Figure 4-16. Outcrop of an 'old' mudstone beds overlain by gravels (near Somerton)

At a point scale, groundwater and surface water level trends were used to infer direction of fluxes. For example, during the two consecutive flood events of 2004, the Carroll Gap reach appeared to be a losing reach during high flow events and switches into a gaining regime during a normal flow events (Figure 4-17), when groundwater head is higher than stream head.

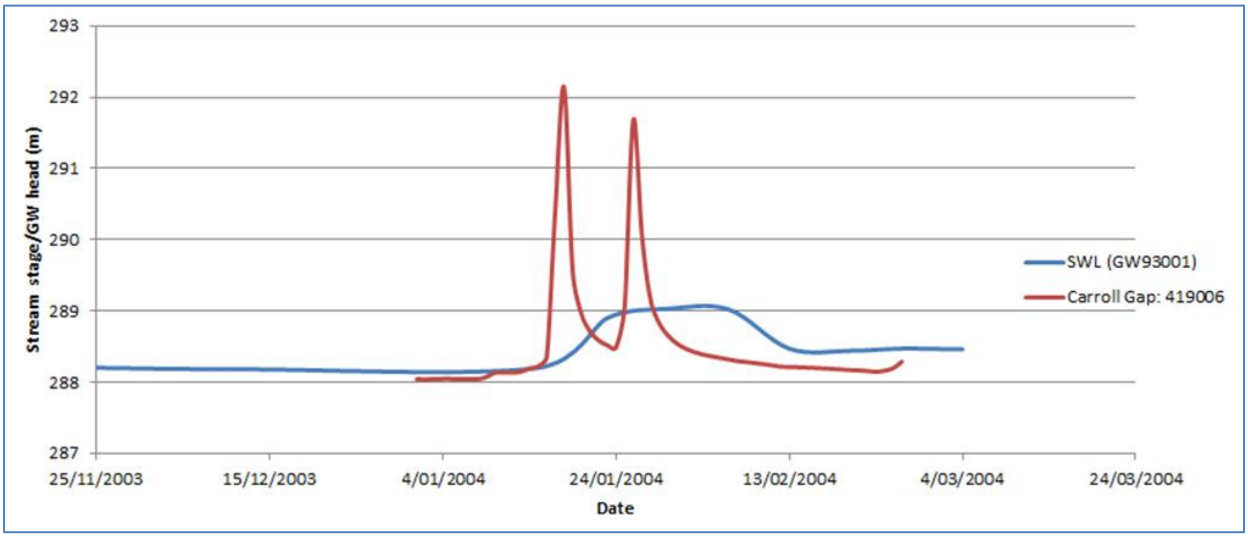


Figure 4-17. Comparison between stream and groundwater heads: Carroll Gap transect

Due to sensor failure, a limited thermal data set is available for analysis. Nevertheless, from the available data set, stream temperatures varied from 9.4 to 14.5°C, while stream stage varied from 0.35 to 0.53 m. Using streambed temperature data measured at depths 40 and 80 cm, near streambed fluxes was calculated. Figure 4-18 shows the best fit match of between simulated and observed sediment temperatures for this site. Based on this match, model-output-estimate of the initial streambed percolation rate varied from 3×10^{-6} to 10^{-5} m/s (Figure 4-18).

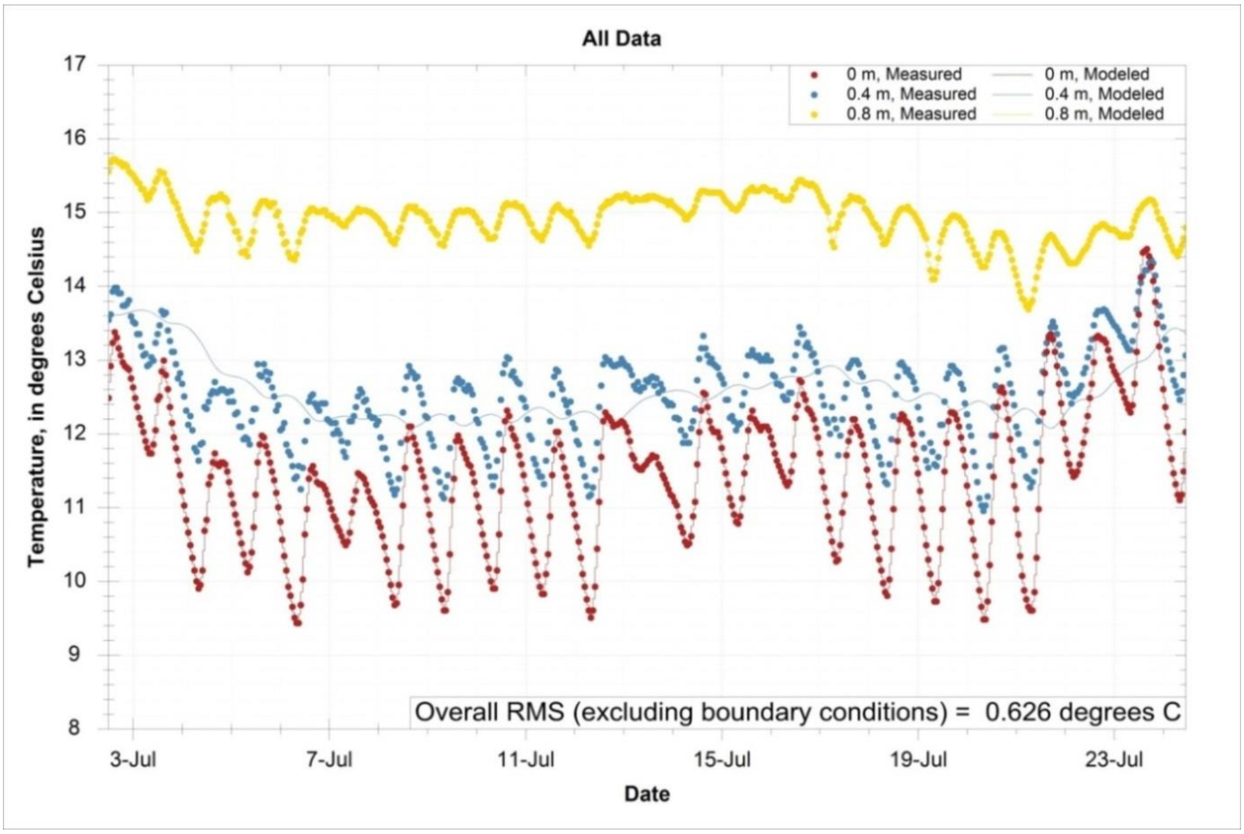


Figure 4-18. Measured Vs Simulated streambed temperature trends at the Carol Gap *transect*

4.3.5.7 Case Study VII: Off Channel Site at Gunnedah

Daily soil temperature data from Gunnedah Climatological Station (GCS, site number: 055024) is available since 1948. After the deployment of soil temperature sensors in May 2007, hourly soil temperature data is available from three depths (10, 20, 50cm). This hourly data set provided an opportunity to estimate the magnitude of the hydraulic conductivity of the local Vertosols and infiltration rates that took place during and after the cyclonic wet spell of 2009 -10. The infiltration rates ranged from 1.20×10^{-6} to 1.25×10^{-6} m/s (Figure 4-19). Again during the simulation, the thermal properties were assumed constant and this may have contributed to discrepancy between measured and soil temperatures for this particular case study.

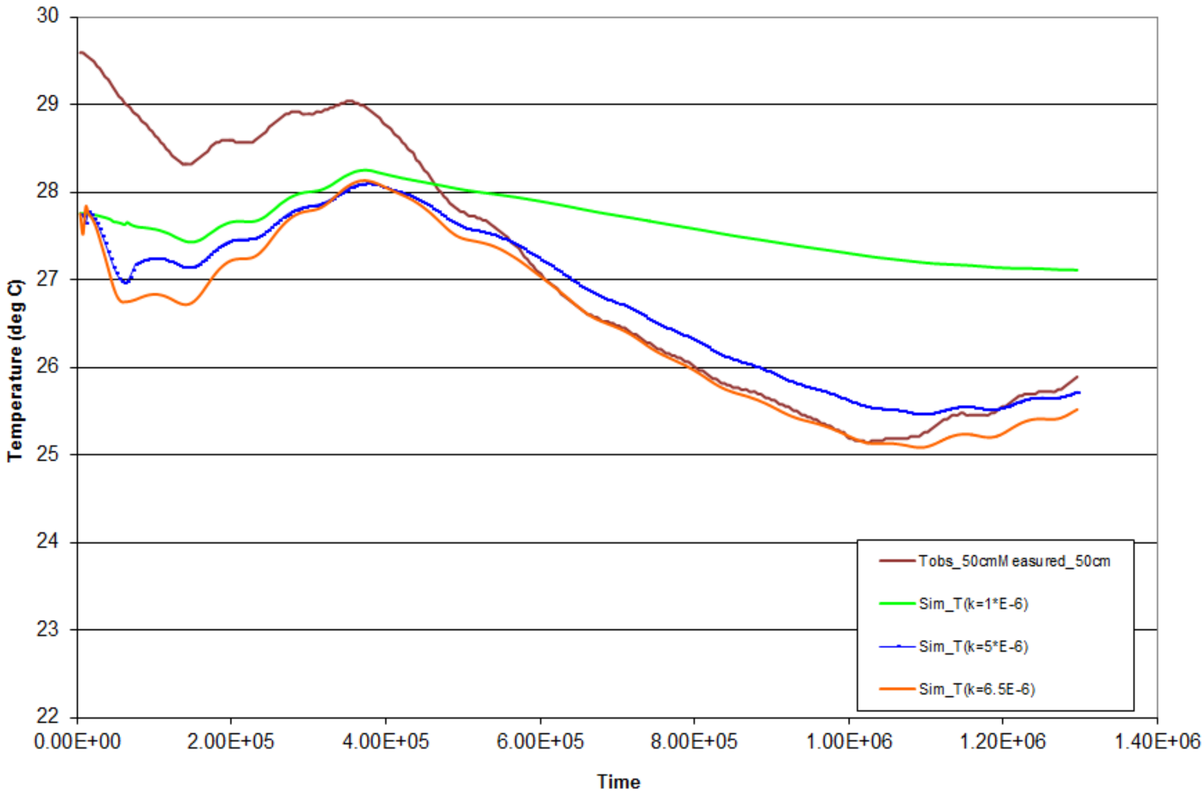


Figure 4-19. Measured and simulated thermographs at depth 50 cm: GCS, site number: 055024

4.3.5.8 Case Study VII: Baldry, Representative of an Ephemeral Stream

The Baldry transect is representative of an ephemeral system. For this site, a simple one-dimensional heat transport model was developed for the P3 site, one of the instrumented reaches at Baldry site. The simulation period covered only a segment of the thermograph where temperature oscillations were detected at depths during stream flow events of June 2010. The simulated and observed thermograph for P3 at 20 cm depth is shown in Figure 4-20(a).

The simulated thermograph mimics the observed thermograph reasonably well during the early period of a flow event in late June 2010. From the 27th June, the differences between measured and observed increased indicating a post flow event, when the streambed starts to dry-up. The predicted infiltration rate ranged from 0 to 6×10^{-6} m/s (Figure 4-20). The estimated K_z is in the same order of magnitude which characterises a decoupled system. It was calculated independently using a sensitivity analysis for discharge into drains (Andersen, 1993).

The analysis of thermal data from the P3 site suggests that the predominant part of the thermograph displays a conductive heat transport mechanism. As the magnitude of seepage rate decreases, simulated temperatures become increasingly more sensitive to variations in streambed thermal properties, such as thermal conductivity of sediments (Constantz, 2008; Hatch et al., 2010).

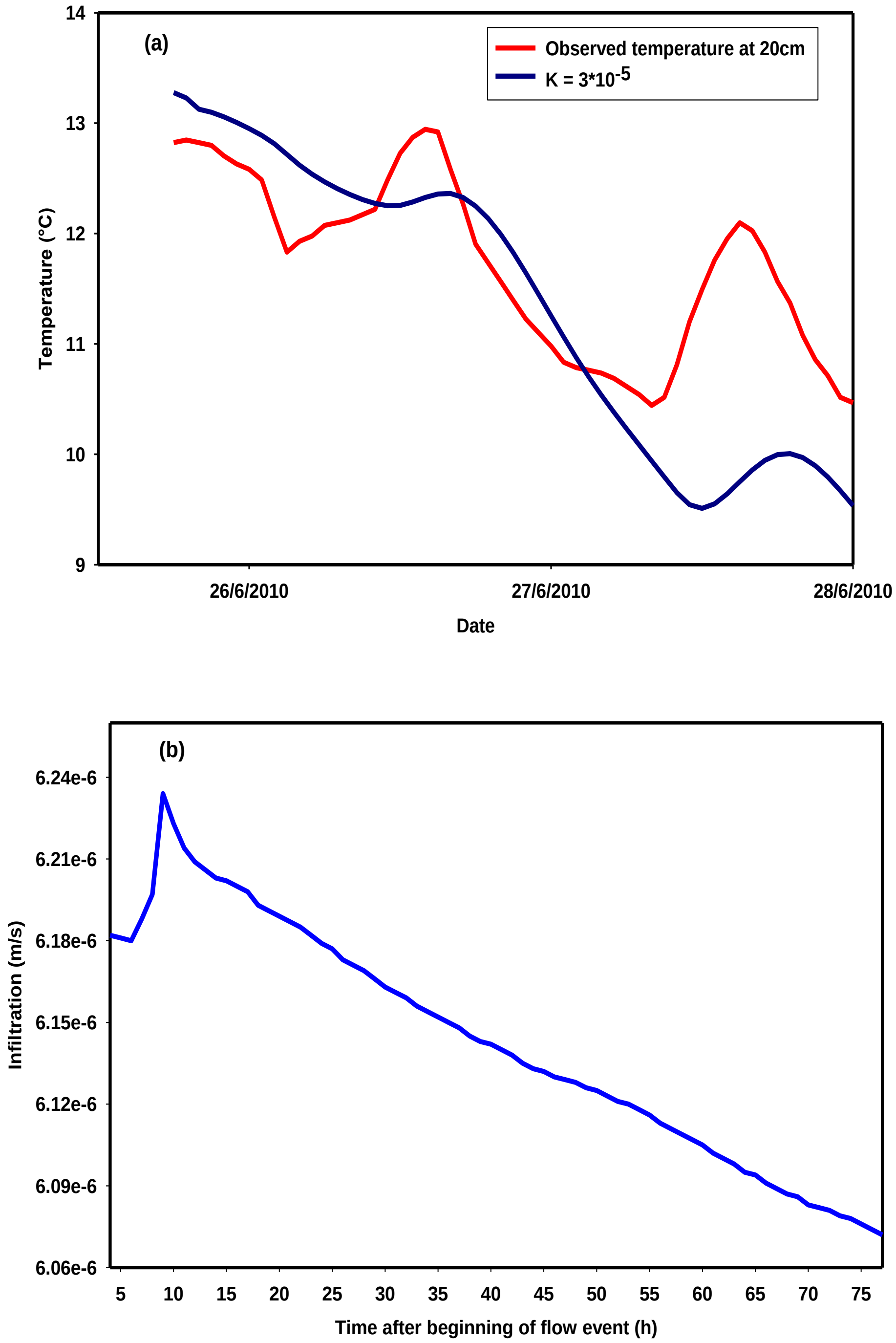


Figure 4-20. (a) Measured and simulated thermographs at depth 20 cm; (b) Simulated infiltration rate

Table 4-4. Calibration performance based on RMS criteria for the different transects of the study area

No.	Type	Transect	Value ^{o}C
1	Regulated	D/S of the Chaffey dam	0.6
2	Regulated	Piallamore	0.012
3	Unregulated	Ballantines	0.048
4	Unregulated	Well 1	0.11
5	Regulated	Carroll Gap	0.31
6	Unregulated	Gunnedah Climatological Station	N/A
7	Unregulated	Baldry	N/A

4.3.6 Management implications

It is well documented in the literature that in base flow dominated river systems (perennial and intermittent streams) groundwater pumping induces reversal of hydraulic gradient and stream depletion. The evaluation of pumping-induced stream depletion is a critical step in the design of catchment scale management plans.

An important characteristic of disconnected streams is that pumping near the stream doesn't induce stream depletion (Winter et al., 1998). From environmental flow management, it would be of a practical importance to determine the critical water table depth, where a stream becomes disconnected for example (Brunner et al., 2010). To my knowledge, there are limited case studies in literature, which assess the impacts of groundwater pumping on ephemeral river systems. For instance (Fox and Durnford, 2003), assessed the impact of pumping on the hyporheic zone. However, their work did not quantify the impact of pumping on stream flow from bores screened below the critical water table depth.

4.3.7 Limitations of heat as environmental tracer

4.3.7.1 Experimental set-up

During the early stage of the experimentation in the Cockburn Valley about 65% of the instrumentation was lost due to flooding that took place between November 2008 and December 2009. One possible option for securing the instrumentation from flood damage is proposed to drill at an angle from the bank of river under the streambed. However, this approach will increase the costs of instrumentation substantially for short term experimentation. The economic viability for a long term monitoring network should be considered in relation to high quality data capture with a minimum risk to flood induced damage.

In addition, hydrological data collected from some of experimental sites (Piallamore and Carroll Gap and mobile zone shown in Figure 5-7) a less reliable on a longer time scale due to sediment build-up in the vicinity of in-stream piezometers. For instance, at the Carol Gap transect, 80 cm sediment built up of coarse gravel was observed during the monitoring period. Thus, the dynamic nature of the upper boundary condition creates additional challenge in data analysis (Chapter 5 of this thesis).

4.3.8 Heat tracing technique is a non-conservative tracer

The transport of heat and solute may be described by the advective-dispersion equation (Eq. 4-6). Despite the similarity between the heat transport equation and the solute transport, there are several differences between heat and solute transport techniques (de Marsily, 1986; Uffink, 1983). The first difference between solute, considered in literature as conservative, and heat transport (non-conservative), is in the diffusive term. On the pore scale, *DT* is significantly larger than the solute dispersion coefficient *D* and induces dampening of high-frequency temperature fluctuations. The second difference lies in the advective term. The linear transport velocity v_T of strictly advective heat transport is smaller than the seepage (groundwater) velocity, v_w:

$$v_T = q\frac{\rho_w C_w}{\rho_b C_b} = \frac{v_w}{R_T} \qquad \text{Eq. 4-6}$$

In which R_T is the dimensionless retardation factor for heat, which can be computed from the volumetric heat capacities of the bulk medium and of water alone:

$$R_T = \frac{v_w}{v_T} = \frac{\rho_b C_b}{n_e \rho_w C_w} \qquad \text{Eq. 4-7}$$

Heat retardation factors between 2.5 and 3.5 were reported for glaciofluvial outwash materials (Giambastiani et al., 2013; Hoehn and Cirpka, 2006; Uffink, 1983). Thus, heat in comparison with other form of environmental tracers (such as Br, Cl, deuterium) is not a conservative tracer (Constantz et al., 2003; Giambastiani et al., 2013).

In addition, the heat tracing technique is less reliable method to determine conduction related thermal parameters in fine grained alluvial sediments. As demonstrated in this study, the Gunnedah and the Baldary sites showed poor fit between observed and simulated streambed temperatures. This may be attributed to variability of thermal properties, which are considered to be diffusive dominated systems. Furthermore, plausible causes for discrepancies between measured and simulated streambed and groundwater temperatures are listed below:

- 1D approximation is valid for shallow depth and short period of simulations. However, with increase in depth, groundwater flow will become horizontal (Domenico and Palciauskas, 1973; Toth, 2009a);

- the assumption of no flow boundary is violated for long period simulations

4.4 Conclusions

In the context of ephemeral and intermittent streams, a number of important physical parameters and physical process, which influence the occurrence and magnitude of river leakages have been highlighted in this chapter. The major conclusions of this study are listed below:

- Apart from the Kootingal reach, which appears to gain on the right hand side the bank and losses on the left hand side of the bank; all of the targeted reaches in this study have been delineated as losing river reaches;
- In the uplands of the Namoi Valley (Cockburn and the regulated section of the Upper Peel Valley), as well as in lower part of the Namoi Valley, the river systems appear to be both losing and disconnected;.
- The physical variables affecting height and lateral extent of groundwater mounding was visualised for a hypothetical ephemeral river system. Physical variables, which have relevance to the development and extent of groundwater mounding in natural channels are: nature of the clogging layer, hydraulic conductivity and aquifer thickness, specific yield, channel width and configuration and depth to water table;
- Apart from few sites, groundwater mounding is not widely observed in the Namoi Valley. Nevertheless, development of groundwater mounding is very sensitive to the hydraulic conductivity;
- A threshold hydraulic conductivity where the pattern of pressure head distributions changes from a bell shaped to a non-bell shaped distribution takes place when K approaches 10^{-8} m/s. The velocity vector is oriented predominantly vertical in the cross-section;

- In intermittent and ephemeral river systems, the relationship between water table and stream seepage can be characterised by two hydrologic regimes (fluctuation and constant flux zone);
- Based on hypothetical models developed for this study, the critical water table depth is about 5 m below channel elevation in some transects. The role of stream depletion caused by groundwater pumping diminishes below the critical water table depth;
- The estimated K_z values at Gunnedah and Baldry sites are in the same order of magnitude; 10^{-6} m/s.
- The thermal method is a robust and affordable method for estimating infiltration/discharge and assessing the effects of dry spells on the stream flow regime;
- Thermal techniques provide a cheaper option to determine infiltration and percolation on a short and long time scale basis;
- In order to increase the level of certainty, we recommend that recharge/discharge estimates based on thermal method should be undertaken in conjunction with other methods such as chemical and isotopic tracers;
- A switching hydrologic regime is considered to prevalent in intermittent rivers and not a characteristic of ephemeral river systems.

4.5 Recommendations

The recommendations of this chapter are:

- As the location of the critical water table depth is an important parameter for environmental flow and water resources management, it should be derived based on field hydrogeological data rather than a virtual model. Geophysical techniques in conjunction with hydrogeological data sets could play an important role in determining the critical water table depth;

- The most pronounced change in temperature was observed within the streambed region and a bulk of the thermistors should be placed in the upper zone of the vertical profile. However, if the ultimate aim of the instrumentation is to estimate recharge, there is a need to install thermistors deep in the vertical profile, extending beyond the regional water table;
- Although the Climatological Station at Gunnedah (40015) is an off-channel site, it was included as part of the experimental sites for demonstration purposes. With a small investment, climatological stations can collect a useful soil thermal data, which can be used for environmental studies (input for climate change related studies).

Chapter 5 Inference of the Dynamic Nature of Streambed Conductance from a thermal streambed data: Cockburn Valley, NSW

'[…] but hydrologic models drift into misuse when they replace thinking […]. An engineer who explains how he sat down on his desk and thoughtfully compared the alternatives will lose out every time to an engineer who presents an elaborate computer program, displays graphic output, and draws a red box around the bottom line.'

L.D. James. (1991); In: D.S. Bowles and P.E. O' Connell, Recent advances in the modelling of hydrologic systems, NATO-ASI series C: Vol. 345, Chapter 26, p.560.

Abstract

The dynamic nature of streambed vertical hydraulic conductivity (VHC) was studied at two sites in the Cockburn River, near Tamworth, New South Wales, Australia. Near streambed fluxes were estimated using the USGS heat transport model, VS2DH, based on thermal data. For modelling purposes, the entire period of observation, extending from July 2008 to May 2012, was split into several segments. In this chapter we analyse in detail a segment thermograph, which represents the two flood events that occurred in late 2008. This illustrates the hydrologic regime prior to the flood events, the effect of floods on the temporal development of de-clogging process and low flow event. During this period, at the pool site, near the Kootingal Bridge the VHC varied from 8×10^{-5} to 10^{-3} m/s and the corresponding fluxes ranged from 1.6×10^{-5} to 4×10^{-4} m/s. In contrast, in an armoured zone, near the Ballantines Bridge, the VHC varied from 8×10^{-5} to 2×10^{-3} m/s, and stream leakage ranged from 9×10^{-5} to 2×10^{-4} m/s. At both locations, a maximum VHC was simulated after flood events, probably induced by the increase in stream temperature and the break-up of the clogging layer by the increase in stream velocity. Available hydrological information suggests that low flow hydrologic regime create a favourable environment for clogging of coarse streambed sediments and disintegration takes place during high and flood events.

5.1 Introduction

Vertical Hydraulic Conductivity (VHC) is an important hydrologic parameter that regulates near streambed water fluxes. Due to physical, biological and chemical factors, VHC is highly variable both in space and time. Thus, the exchange processes between surface water and groundwater vary in both space and time (Brunke and Gonser, 1997; Woessner, 2000). Temporal fluctuations may be caused by variations in focused and distributed recharge, whereas spatial variations are caused by streambed heterogeneity and associated hydraulic conductivity (Fleckenstein et al., 2006 ; Kalbus et al., 2009; Vogt et al., 2012).

Apart from limited studies carried overseas, a review of the available literature suggests that there is a lack of knowledge on the dynamic nature of the clogging layer in natural river systems. Although there are a number of research work carried out in a controlled laboratory environment such as flumes, the outputs from these experimentation, may not necessarily replicate hydrological processes, occurring in a natural world. In one of earliest studies on seepage Bouwer (1969); Bouwer (1978); Bouwer (1982), provides a theoretical presentation of seepage from open channels with clogged, wetted perimeters. Based on this methodology, an Excel based spread sheet for estimating stream leakage is included the Appendix C section of this thesis.

In most regional groundwater models, due to paucity of data, VHC is assumed to be constant. However, this assumption may not be valid for small scale investigations and groundwater pollution and remediation studies. Therefore, we concentrate in this chapter on temporal variability of VHC. In most mechanistic models, the exchange of water between an aquifer system and a stream/creek is expressed by the equation:

$$q = \frac{(VHC) * LW}{M} * (H_{aq} - H_{ck}) \qquad \text{Eq. 5-1}$$

Where q is flux between the aquifer and the stream, VHC is the vertical hydraulic conductivity of the stream/creek bottom sediments, L is the stream width, W is the width of the stream, M is the stream bed thickness, H_{aq} is the hydraulic head in the aquifer and H_{ck} the water level (stage) in the creek.

From this it can be observed that the water fluxes between surface water and groundwater systems are a function of the difference between river stage and aquifer head. Down welling will occur whenever the river stage exceeds the groundwater head, but the rate depends on the magnitude of the difference, the conductivity of riverbed deposits, and the saturated area of the channel. In addition, streambed exchanges (hyporheic exchange) as discussed by (Kaser et al., 2009) can be grouped into five distinct classes. This issue is presented in detail in Chapter 6 of this thesis.

In a natural system, however, settling and straining of suspended and bed load sediment at the riverbed may cause a substantial reduction of the conductivity of the outermost layer. These processes are usually referred to as clogging (Joppen et al., 1992; Lilse, 1989; Schalchli, 1993). One of the hydraulic properties of a clogged layer is expressed in terms of streambed conductance, which is defined as VHC divided by thickness of the layer (VHC/M). This important parameter is difficult to estimate due to a lack of knowledge of VHC and thickness of the clogging layer (Dahl et al., 2007).

Another major assumption made in regional groundwater models is that a linear relationship between flux and Δh exist. This is often too simplistic and does not take into account the decreased hydraulic resistance as the stream volume and velocity increases (Eertwegh, 2004; Rushton and Tomlinson, 1979), due to de-clogging of the upper streambed layer. Bedload movement, which occurs during high runoff and flood events, facilitates the de-clogging process. In addition, stream reaches with upwelling 'groundwater', tend to maintain hydraulic connectivity between streambed and groundwater, due to upward hydraulic forces reducing siltation (Hatch et al., 2010; Schalchli, 1993).

Natural river channels are dynamic and depending on the scale of time, deposition and erosion may occur on different reaches of a river system. For illustration purposes, we provide river reaches from the Peel Catchment, where deposition and aggradation has taken place on different reached of the river, since monitoring started. At the Carroll Gap (Gauging Station: 419006) reach, the lowest recorded stream stage is lower than the current streambed elevation; based on this historical information, one could infer that aggradation has taken place (Figure 5-1 and Figure 5-2). In contrast, at the Gauging Station, the river has incised its channel in time. The consequence of creating an incised channel is associated with accelerated streambank erosion, land loss, aquatic habitat loss, lowering of water tables, land productivity reduction and downstream sedimentation (Rosgen, 1997).

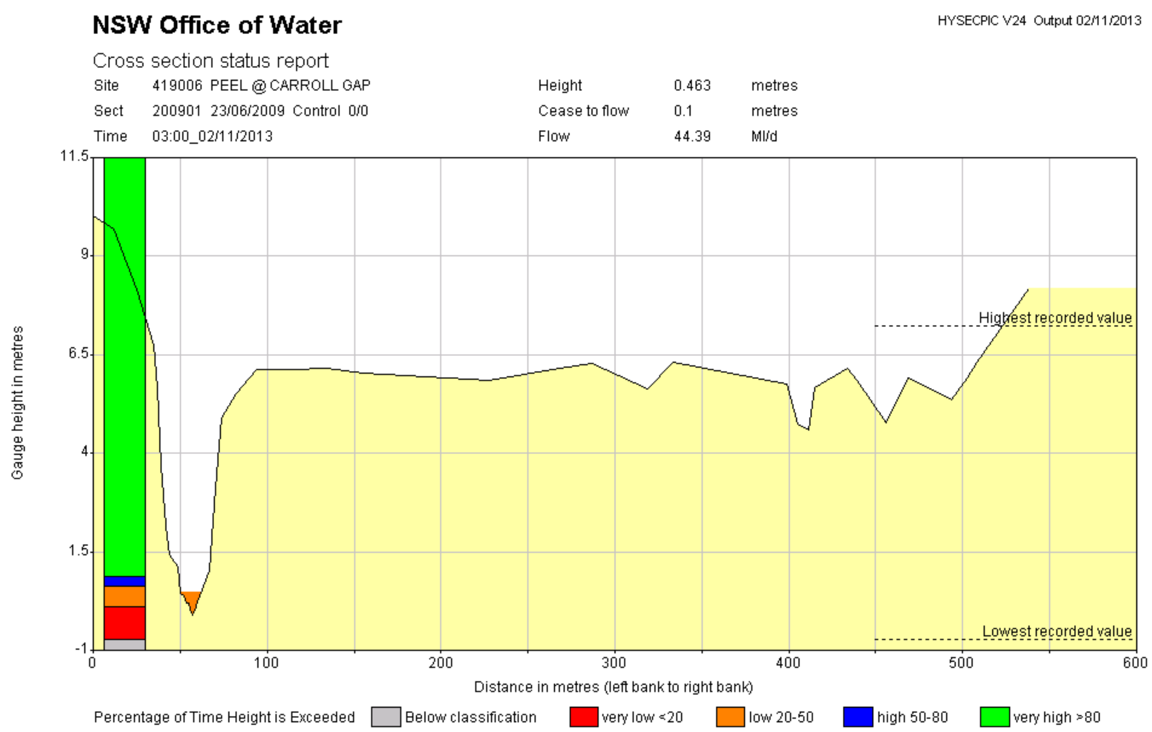


Figure 5-1. A representative of aggradational reach in the Peel Valley: Carroll Gap (419006)

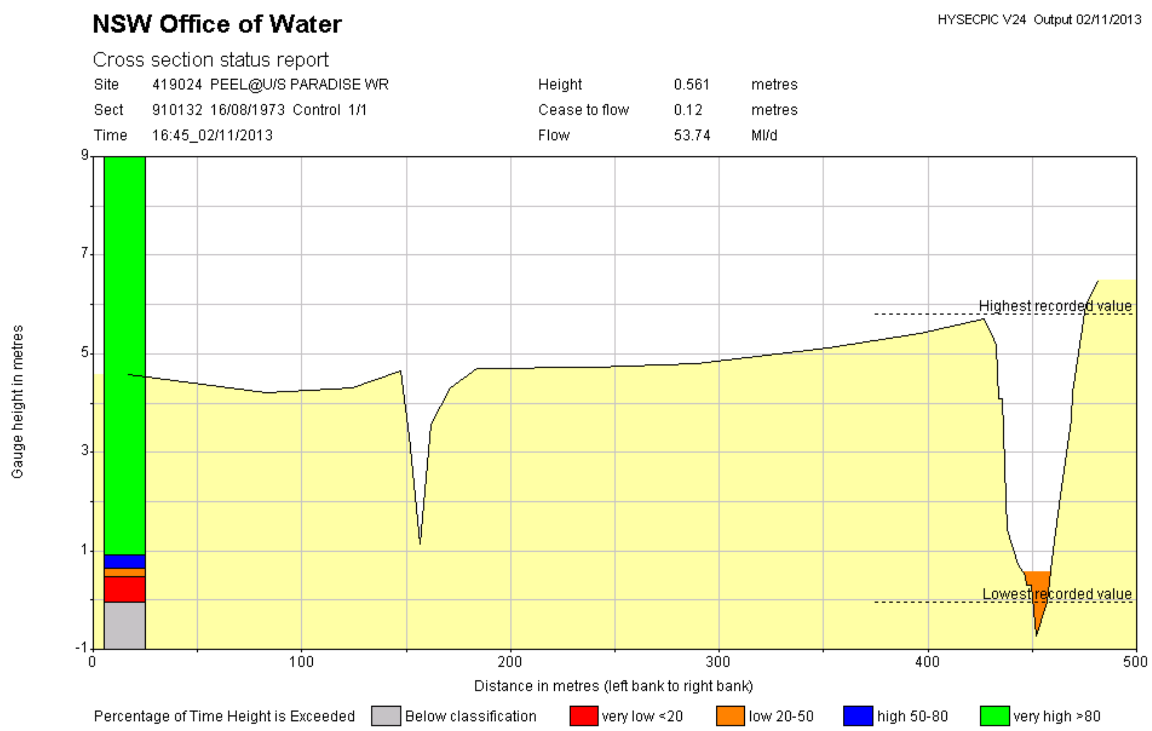


Figure 5-2. A representative of a degradational reach in the Peel Valley: Paradise Weir (419024)

During the last decade, due to the availability of cheap and reliable thermistors, several investigators have successfully used heat as an environmental tracer for estimating near streambed fluxes and VHC (Anderson, 2005; Bartolino and Niswonger, 1999b; Constantz et al., 2007; Cox et al., 2007; Hatch et al., 2010; Mutiti and Levy, 2010). Most of these studies covered a short time frame. In addition, few of these investigators targeted the dynamic nature of VHC explicitly. In this study, we have used extensive hydrological data collected over 48 months, to understand the spatial and temporal variability of VHC in the Cockburn Valley. The data collected covers a range of hydrologic regimes including both low flow and flood events.

I assume that the dynamic nature of VHC is linked to the hydrologic regime of the river. During low flow hydrologic regimes the critical tractive force responsible for mobilizing riverbed sediments is low. Consequently, there is a tendency for the reaches to deposit rather than transport sediments. In contrast, during high river flow the transport of sediments may be prevalent (Figure 5-2).

The main objectives of this chapter are to: (1) Study the variability of a vertical hydraulic conductivity in space and time, using hydrological data sets collected during flooding events of 2008; (2) Present variability and comparison of seepage fluxes for different segments of the streambed thermographs; and (3) Infer dry spells or 'disconnection' of streams using streambed temperature measurements at different depths. This particular application is considered to have relevance to the environmental flow management of ephemeral river systems in the MDB.

5.2 Study Site

The study site is located in the Cockburn Valley, NSW. The Cockburn River joins the Peel River southeast of Tamworth. In the catchment, the alluvial aquifer of the valley is restricted to a narrow zone between Nemingha and Ballantine's Bridge, a length of approximately 15 km of varying width between 1-3 km and a gradient of 40 m in 15 km (Calaitzis and Leitch, 1994). Generally the top 6 metres of the alluvium consist of finer sediments and the main water bearing zones occur below this depth. There are a large number of shallow wells and bores in the alluvium, which are generally used for irrigation. The fractured rock aquifer systems are characterised by Carboniferous meta-sediments and granites of the New England Batolith (Figure 5-3.). Thus the study area is dominated by a hard rock geology rather than unconsolidated sediments, which becomes prevalent downstream of Tamworth. Due to the impermeable nature of surface geology and steep topography, the Cockburn River exhibits one of the highest specific discharges in NSW.

Due to lack of data, there is an ambiguous relation between occurrence, geological structure and lithology in the Cockburn Valley. Further work is required in clarifying the role of fracture controlled groundwater discharge.

The average annual rainfall decreases with decreasing altitude from the highest point in the Great Dividing Range (1440 m) to Tamworth (280 m) and rainfall ranges from 800 to 650 mm/year.

The study area is an ideal location to study surface groundwater connectivity because of the relatively high density of gauging stations and hydro-geomorphological factors. Three permanent (419016, 419056 and 419099) and two temporary gauging stations (TGS) are operated by the NSW Office of Water. The gauging station at Mulla Mulla Crossing has operated since 1936. At this particular gauging station, the catchment area is about 900 km2 and elevation is 441 mAHD.

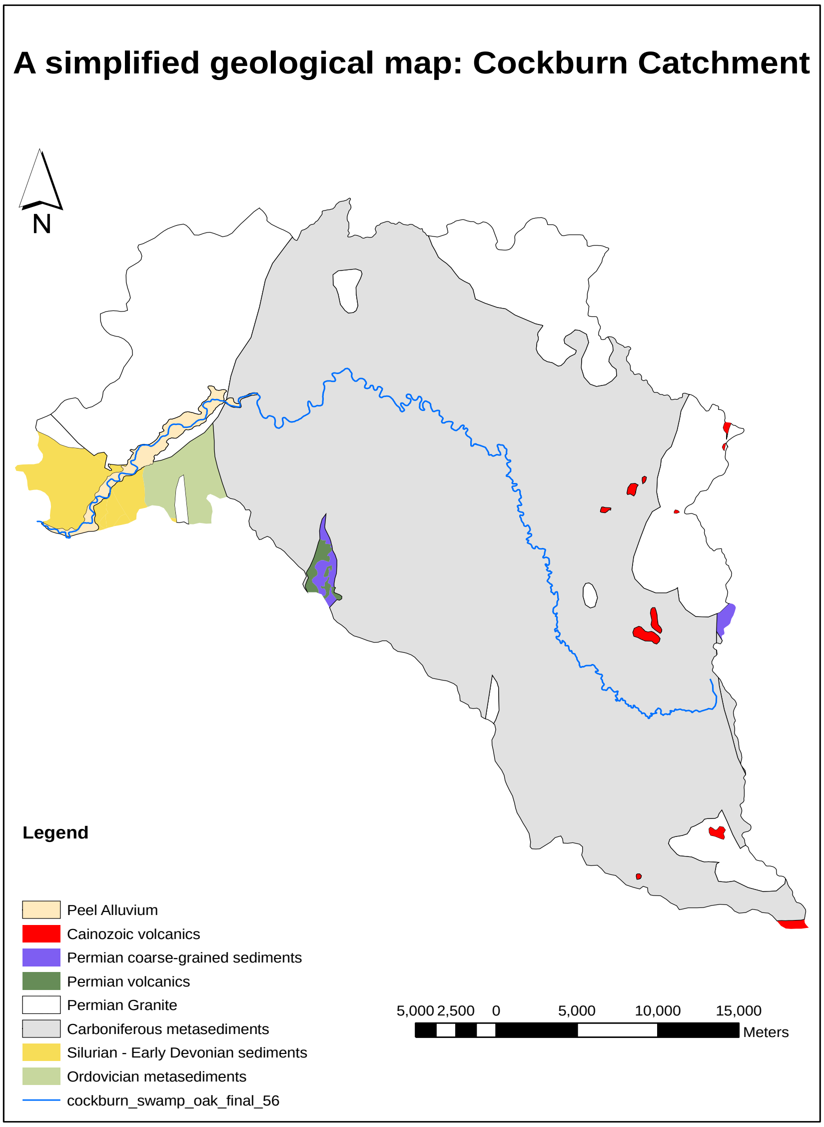


Figure 5-3. A simplified geological map for the Cockburn Valley

In addition, there are 7 NSW Office of Water monitoring bores (GW93036-40) equipped with loggers, which were supposed to measure groundwater levels and temperatures at an hourly interval. Two additional wells were equipped with loggers as part of this study (Figure 5-4).

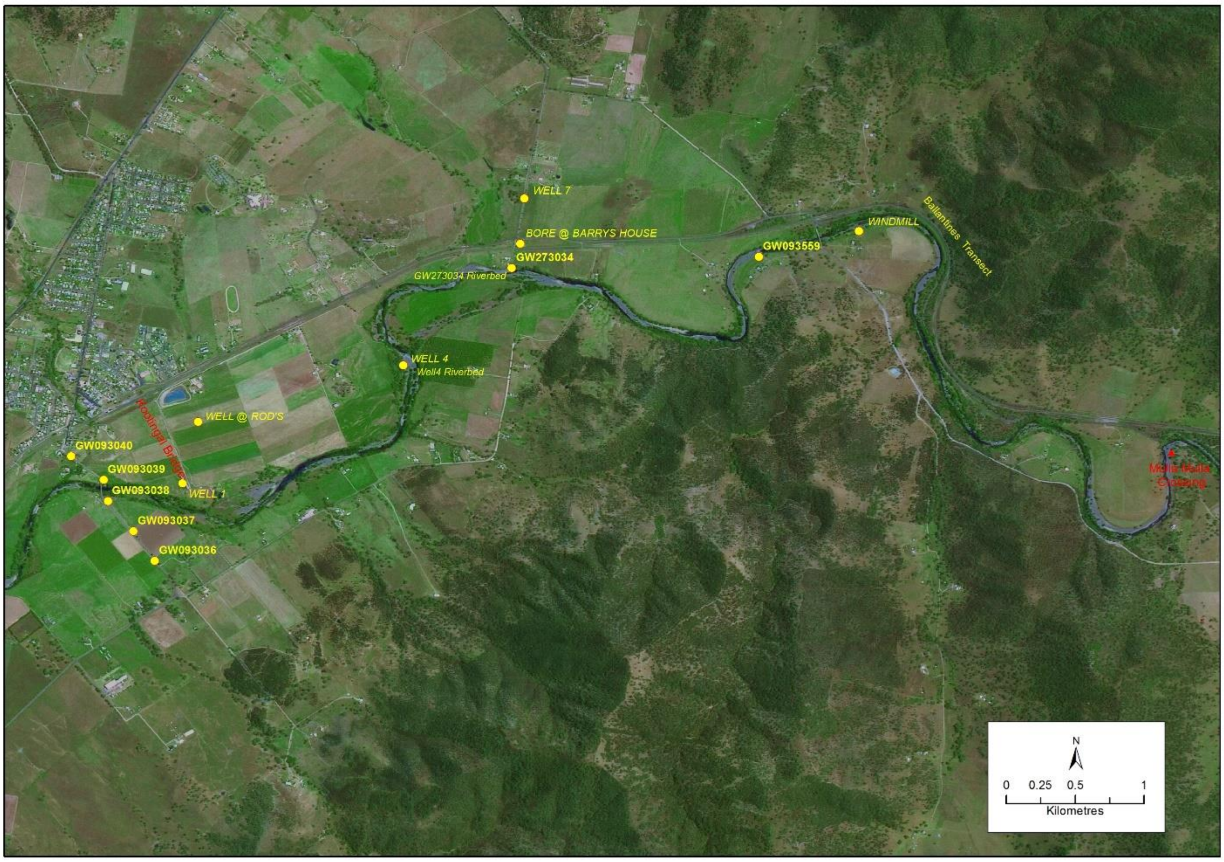


Figure 5-4. Locations of the monitoring bores in the Upper Cockburn Valley transect

5.3 Methods

5.3.1 Instrumentation the designated transects in the Cockburn Valley

At selected reaches of the Cockburn River (Ballantines and Kootingal transects (Figure 5-4), 37 mm diameter galvanized pipes were hand driven to 1 m depth below streambed elevation (Figure 5-5). In this case study, the Kootingal transect is representative of a pool zone; while the Ballantines transect is a representative of riffle/pool zone with an armoured geomorphology.

HOBO® Temperature/Light Data Loggers (at 25 °C, accuracy: ±0.47 °C; resolution: 0.1 °C) were suspended on a string and inserted below the riverbed to the required depths inside the piezometers. As the riverbed sediments predominantly consist of gravels, cobbles and boulders, construction of in-stream piezometers of 30 mm diameter by hand was extremely difficult and time consuming (Figure 5-5(B)).

A

B

Figure 5-5. A galvanised pipe used for construction of an in-stream piezometer was designed and constructed in a local workshop at Gunnedah (A) and a fence post pusher was used to install piezometers in transects of the study area (B).

5.3.2 Heat Transport Modelling

After review of available heat transport models, the USGS heat energy and water flow model VS2DH (Healy and Ronan, 1996a) and its graphical user interface VS2DI (Hsieh et al., 2000b) were considered to be suitable for the hydro-geomorphological environment of the study area for the following main reasons (additional background info is provided in Chapter 3 of this thesis):

- VS2DH package has been applied extensively for arid and semi-arid environments in the US and other parts of the world. A number of investigators have used VS2DI to estimate near streambed water fluxes (Constantz, 2008; Constantz et al., 1994; Su et al., 2004).
- VS2DH solves heat and water transport in unsaturated and saturated model domains, which is a public domain package and user friendly (See Chapter 3).
- 1DTemPro, a user friendly interface for VS2DH was used to estimate near streambed hydraulic conductivities and fluxes.

Given the limited soil and groundwater temperature data available for the study area, the instrumented sites in the Cockburn valley were modelled in one dimension using 1 column and several rows of different grid spacing. The total domain length was 15 m (Figure 5-6), representing the maximum thickness of the alluvial deposit in the Valley and grid spacing varied from 0.1 at surface to 0.25 m at depth.

Stream stage and temperature data for the Mulla Mulla and the Kootingal gauging stations are available since April 2008 and this data set represented the upper boundary of the model domain. The specified head condition allows a flux of water (recharge or discharge) to cross the water table during modelling. No-flow conditions were assigned to the lateral boundaries. As field data on hydrothermic properties were not measured in the field, these properties were derived from literature (Table 5-1).

At this particular site, the river gains on the RHS of the bank and loses on the LHS of the bank. Due to complexity of the boundary conditions, a 2D model was not deemed to be appropriate. Having this in mind, the modelling effort was kept as simple as possible and a one-dimensional model was constructed to estimate VHC and fluxes.

Table 5-1. Summary of hydrothermic properties used in VS2DH simulations

Parameter	Value	Source
Hydraulic conductivity (kz)	Calculated	Site specific
Porosity	0.37-377 (m3/m3)	Site specific
Heat capacity of dry solids (Cs)	1.2×10^6 - 2.18×10^6 (J/m3 °C)	Niswonger & Prudic (2003)
Heat capacity of water (Cw)	4.18×10^6 J/(m3 0C)	Su et al. (2004)
Thermal conductivity (kt)	1 W/(m3 °C)	Su et al. (2004)
Longitudinal thermal dispersivity (al)	0.5	Niswonger & Prudic (2003)
Traverse thermal dispersivity (at)	0.01 to 0.1 (m)	Niswonger & Prudic (2003
Anisotropy ratio (Kv/Kh)	1	Site specific

From my experience in this study, 1D model simulations appear to be more unstable. Due to numerical oscillation and numerical dispersion, the clogging layer was not modelled explicitly for certain combinations of k_z and grid size, during the initial stage of simulations. Instead, the profile was considered homogenous with a total thickness of 15 m, overlain by a clogging layer of 0.25m. Based on this, the VHC of the clogging layer was inferred from the depth-weight hydraulic conductivity equation (Bouwer, 1978).

$$k_z = \frac{z}{\sum_{i=1}^{n} \frac{d_i}{k_i}} \qquad \text{Eq. 5-2}$$

Where:

k_z = vertical average hydraulic conductivity (VHC) (m/s);

k_i = vertical hydraulic conductivity of layer *i*

z = total thickness of the alluvium (m)

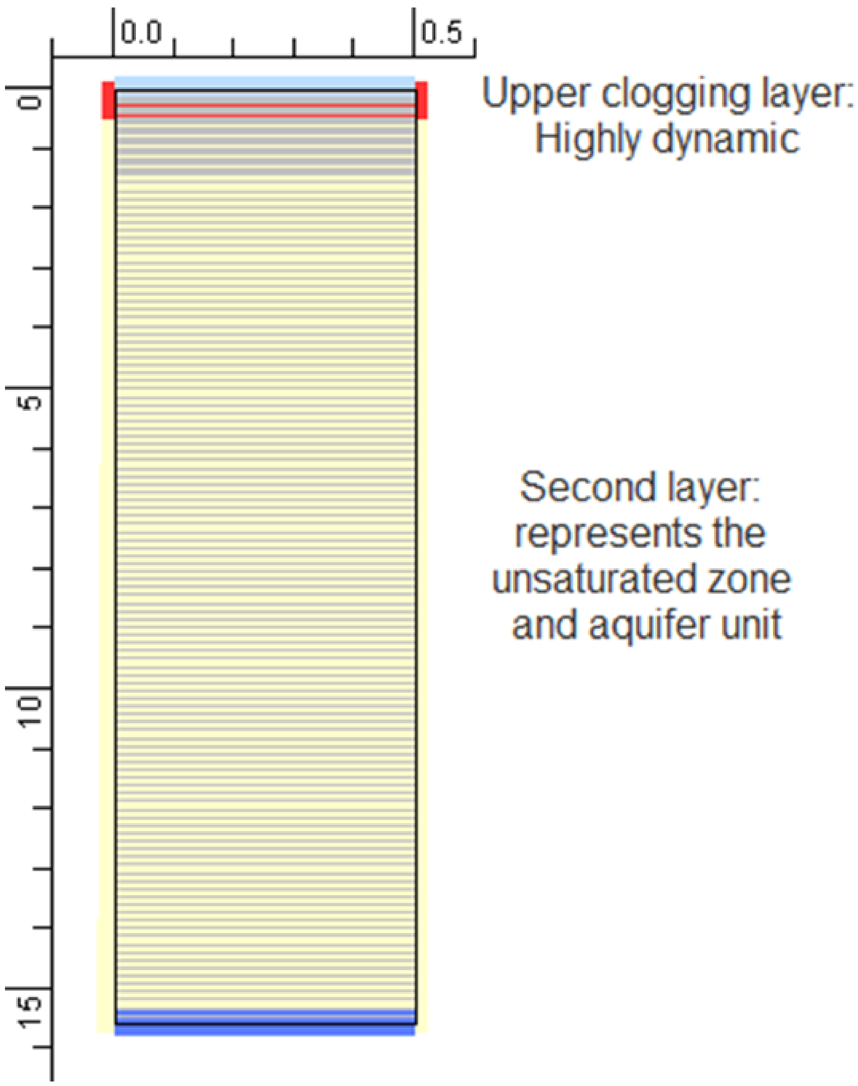


Figure 5-6. A clogging layer represented in a simple 1D

At a later stage, the model domain was extended artificially to 30m length to minimise the effects of a lower boundary on model simulations. This modelling exercise encountered less instability (5.4.1.1.2). It is important to keep in mind that the selection of boundary conditions has significant influence on the water balance components (Berhane, 2006).

5.3.3 Automatic Calibration

A better fit between measured and simulated thermographs was achieved by using a combination of a manual and automatic calibration approach. As manual calibration is very tedious and time consuming, an automatic calibration technique such as PEST was used in this chapter. PEST is based on the Gauss-Marquardt-Levenberg optimization algorithm (Doherty, 2004). During calibration, PEST automatically reruns the VS2DH models until the objective function is minimised in a weighted least-squares sense using the measured and simulated temperatures. The objective function is presented by the following equation:

$$S(\varepsilon) = \sum_{i=1}^{n} \left[T_{oi} - f(x_i, \varepsilon)\right]^2 \qquad \text{Eq. 5-3}$$

Where, ε is the set of parameters to be optimised; T_{oi} is the field observation; $[T_{oi} – f(x_i,\varepsilon)]$ is the difference between observed and model generated result; and, $S(\varepsilon)$ is the objective function. An automated calibration process using PEST was used to determine the set of parameters generating the lowest value of the objective function.

5.4 Results

5.4.1 River Reaches /Geomorphology

In this section the geomorphology of the study area has been analysed from previous reports and field observations during the experimentation phase of the project. For instance, installation of in-stream by hand was a daunting task in the upper part of the Cockburn Valley, due the presence an indurated armoured layer, below 40 cm. In addition, establishment of additional monitoring bores in valley had its own set of challenges and was very expensive.

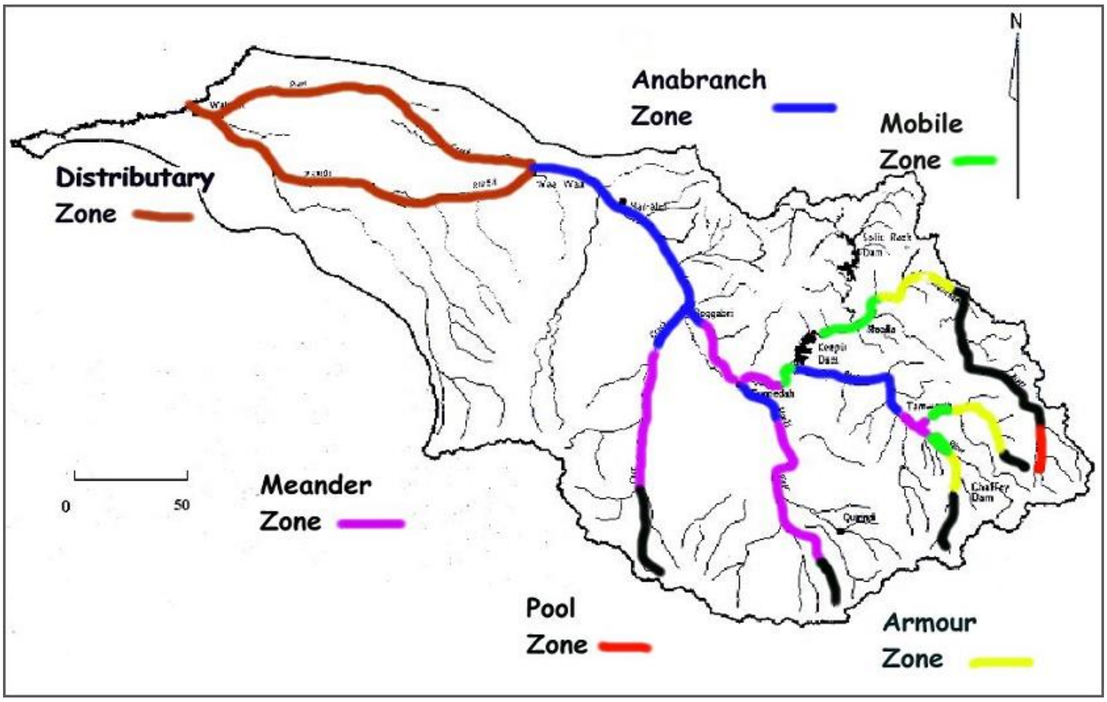


Figure 5-7. River zones in the Namoi Valley catchment (after Thomas, 1988)

Based on a geomorphological survey of the Namoi Valley, (Thoms et al., 1999) has identified seven bedforms or river reach types (constrained, pool, armoured/riffle, mobile, meander anabranch and distributary in (Figure 5-7). This method of classification is in line with my field observations, with the exception that the size of some features may have altered after 1998. The first type of river zone occurs in the upper reaches of the sub-catchments, including the Cockburn Valley. The location of the zones reflects the variable control of the discharge of water and sediment in relation to catchment size and geological influences on the nature of the valley. Similarly, the extent of each zone varies according to the overall geomorphology of the region.

Streambed sediments in the upper part of the study area consist predominantly of boulders, cobbles, gravels and silt. Bed armouring was observed in the riffle sections between the Ballantines and Well 4 transects. In the lower part of the study area, the sediment size decreases, the beds consist of poorly sorted cobbles, gravels, with variable amounts of clay and silt material. Due to streambed metamorphosis, the hydraulic properties at streambed scale are highly variable both in space and time.

In the context of this project, human induced changes play a role in surface and connectivity study. For instance, sand and gravel extraction on the river has contributed significantly to the degradation of the river. As reported by Lagasse and Simons (1976), "the removal of gravel from the river can seriously impact river morphology and stability by removing significant quantities of the coarser sediments from the river. This coarser fraction, particular gravel, has a tendency through hydraulic sorting to armour the riverbed, thereby retarding and arresting excessive scour, stabilising banks and bars, and preventing excessive sediment movement. Gravel armoured sandbars serve as semi-permanent controls that define river form. Removal of the gravel armour can lead to erosion and loss of this control." For example, wooden supports for the old Kootingal Bridge, which were originally cut off at riverbed level, are now up to a metre above the riverbed (Figure 5-5).

In the Cockburn River, the excessive mining in the past has caused bed lowering and degradation. The lowered baseline of the riverbed caused by degradation increased the slope of the mouth of the tributaries along the reach. This increased slope results in upstream progressing degradation of the tributaries (Galay, 1983).

5.4.1.1 Reach I: Water hole/ Pool Zone at the Kootingal transect

Pool zones in the study area are characterized using long pools separated by short channel constrictions. The pools form upstream of these channels constrictions and are the dominant morphological feature in the zones. Channel constrictions are generally associated with major bedrock bars that extended across the channel or substantial localized gravel deposits that act as a riffle area (Lagasse and Simons, 1976). The Kootingal pool is the largest pool system in the Cockburn River; this system is partly fed through flow system. It dries up only during extreme drought events as observed in 2006, just prior to field experimentation started in the Cockburn Valley. In order to enhance the existing conceptual model, the study site was surveyed in detail in May 2009, the survey results in combination with the available hydrological information assisted to get additional insights on the through flow system in the catchment.

5.4.1.1.1 Groundwater level and temperature trends

Groundwater thermographs for bores (GW93036, GW93039, GW93040, GW93037) screened in gravels at different depths (8.5-11.5, 12.5-15.5,13.5-16.5,17.5-20.5 mbgl) respectively and different horizontal distances from the bank of the Cockburn River were decomposed into Fourier components, to determine the phase and amplitude of the annual frequency components for bores with long term data (Berhane et al., 2008). The average groundwater temperatures ranged from 18.9 °C to 19.6 °C. The surface temperature propagates with decreasing amplitude into the groundwater. The highest amplitude of fluctuation was observed in GW93040 and lowest in GW93036. Unfortunately, the NSW Office of Water has ceased to collect groundwater temperature data starting from September 2008 and further analysis on groundwater temperature trends is not possible.

Interestingly, this particular transects gains water on the right hand side of the bank and loses water on the left hand side of the bank (Figure 4:9). The available data suggest that the groundwater table is far below the channel level for GW93036-38, indicating a losing stream condition. In contrast, water levels in bore GW93040 are higher than the streambed elevation for some flow periods, indicating the groundwater system may be contributing to stream discharge after runoff and flood events. However, the in-stream piezometer located at the RHS didn't suggest the stream is gaining at this transect. Probably, the groundwater seepage is insignificant when compared to stream discharge.

The conceptual model shown in Figure 5-8 was corroborated by groundwater temperature measurements in piezometers (GW93036-40). During flood events, a slowly changing groundwater temperature was disrupted by an advective heat transport mechanism, which is very noisy compared with the slowly changing temperature trend in the gaining section of the groundwater temperature profile (Figure 5-9).

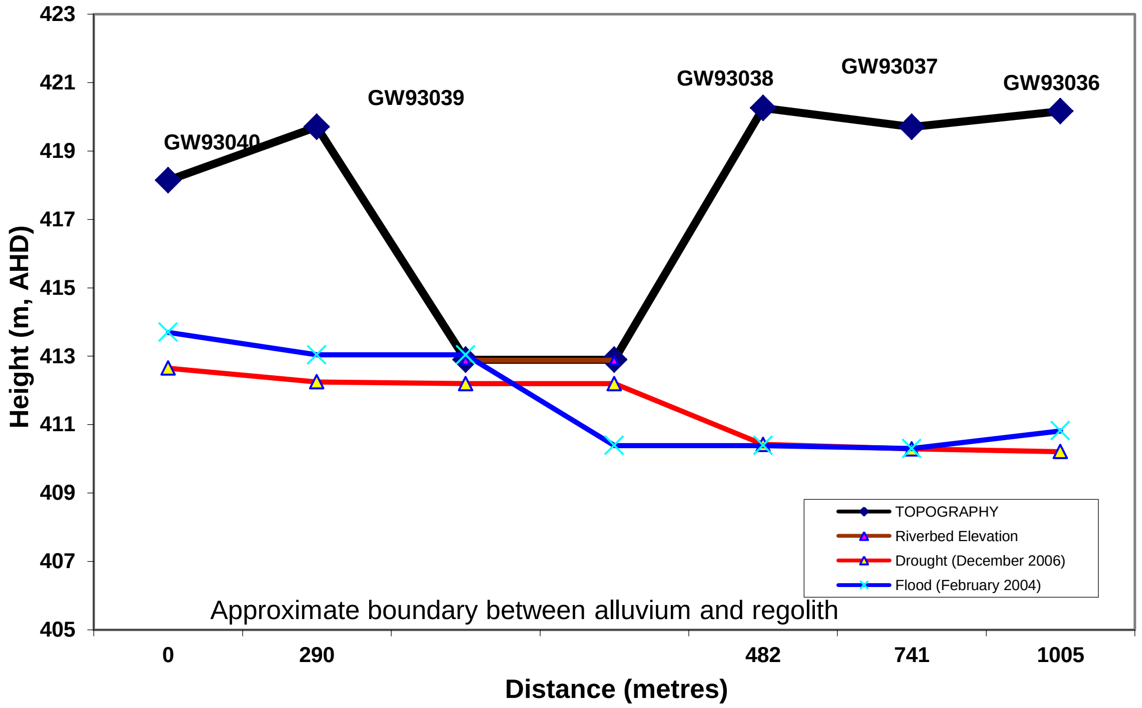


Figure 5-8. Groundwater dynamics at the Kootingal transect (direction of stream flow is towards the reader).

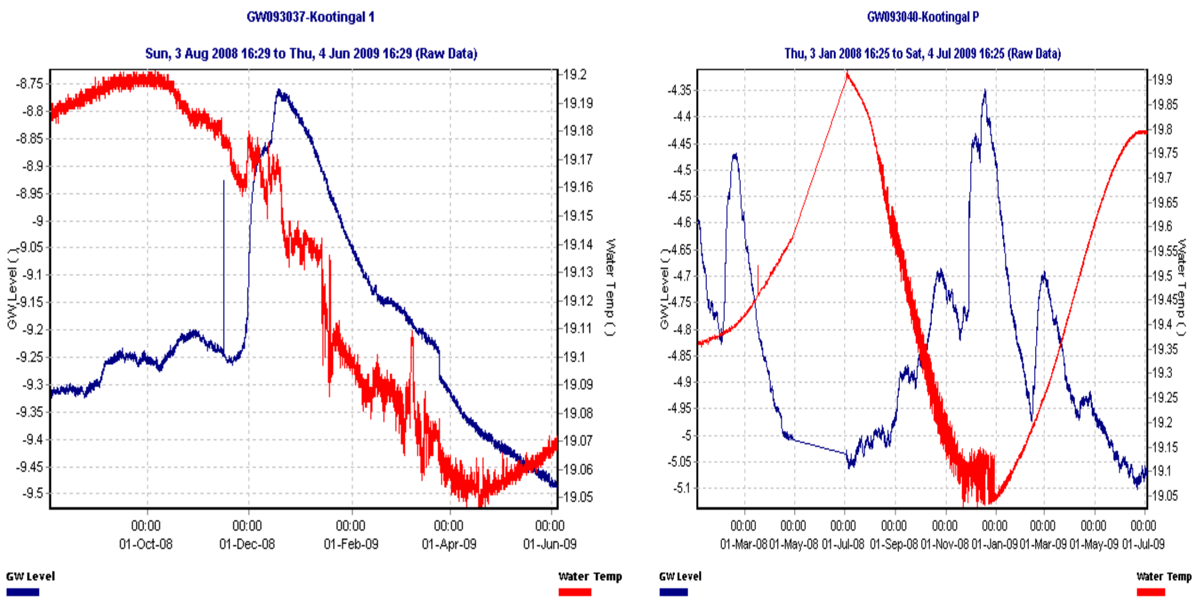


Figure 5-9. Contrasting groundwater temperature responses in losing (GW93037) and gaining sections (GW93040)

The Ballantines transect was surveyed in detail using a differential GPS in 2010. During this survey, piezometer GW93559, located on the left hand side of the bank and two privately owned wells on the right hand side of the bank were included. The available hydrological data suggest that the groundwater is far below the channel level for GW93038, indicating a losing stream condition (Figure 5-10).

5.4.1.1.2 Streambed thermograph analysis

As part of this study, the existing monitoring infrastructure in the Cockburn Valley was extended and the monitoring at the Kootingal transect extended for more than three years (Figure 5-11). During this period (14/07/2008 to 12/05/2012) events consisting of single, multiple and complex events were observed. From the start of analysis of streambed thermograph data, it was obvious that variability of streambed hydraulic conductivity each of the streamflow event may not necessarily be captured by a simple deterministic model for the entire period of observation.

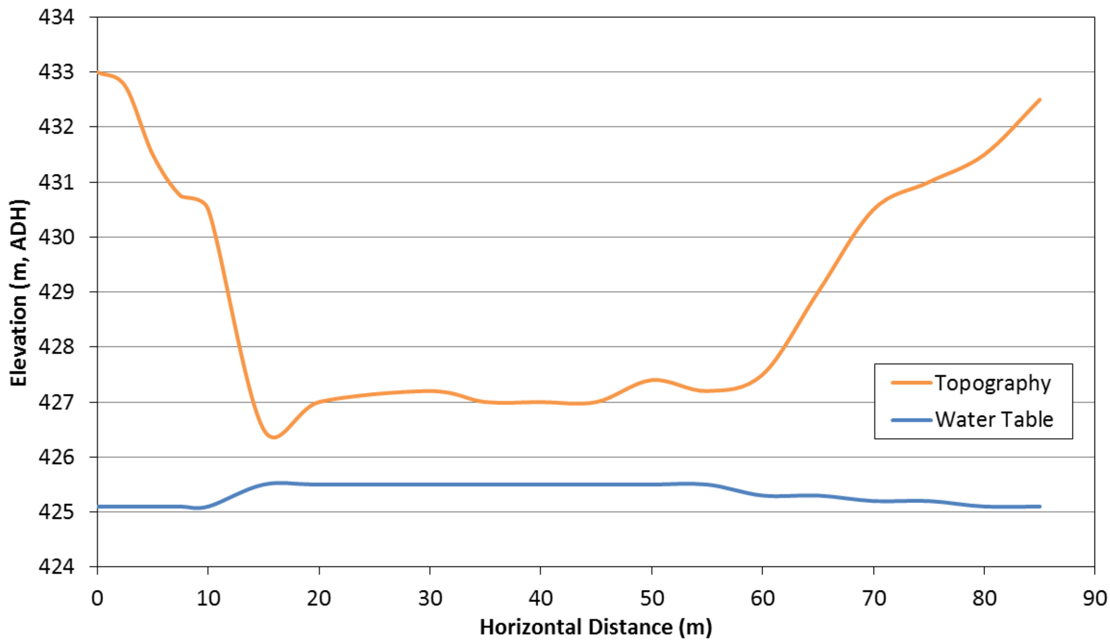


Figure 5-10. Relationship between water table and streambed elevation at the Ballantines transect.

During the entire observation period, stream temperatures varied from 7.6 in winter to 29.7 °C in summer, while stream stage varied from 0 to 6.0 m. For the same period, streambed temperatures measured at depths 40 and 80 cm below the channel varied from 8.49 to 29.97 °C and from 8.97 to 29.77 °C respectively (Figure 5-12).

5.4.1.1.3 Best-Fit Hydraulic Conductivity for the entire period (First Stage)

Analysing thermal data for the entire period of observation may provide hydrological information on ‘average’ seepage rates and ‘average’ hydraulic conductivities. At this stage, the significance of an average value for a small scale investigation and ephemeral/intermittent river systems is not clear. Having this issue in mind, the entire period of observation was divided into several segments, which are considered to be representative of hydrologic regimes (hydrology, biology, chemistry and other an accounted physical and anthropogenic factors).

However, prior to providing event based thermal analyses, streambed thermal conditions at the Kootingal Bridge were analysed in three stages. During the first stage, best fit hydraulic conductivity was obtained for the period from to 29/07/2009 to 12/05/2012, where streambed temperature measurements are available at 40 and 80 cm depths (Figure 5-12). For this particular exercise, 1D modelling endeavour was facilitated by 1DTempro. Observed and best-fit streambed temperatures for an average hydraulic conductivity of 6×10^{-5} m/s are shown in Figure 5-11.

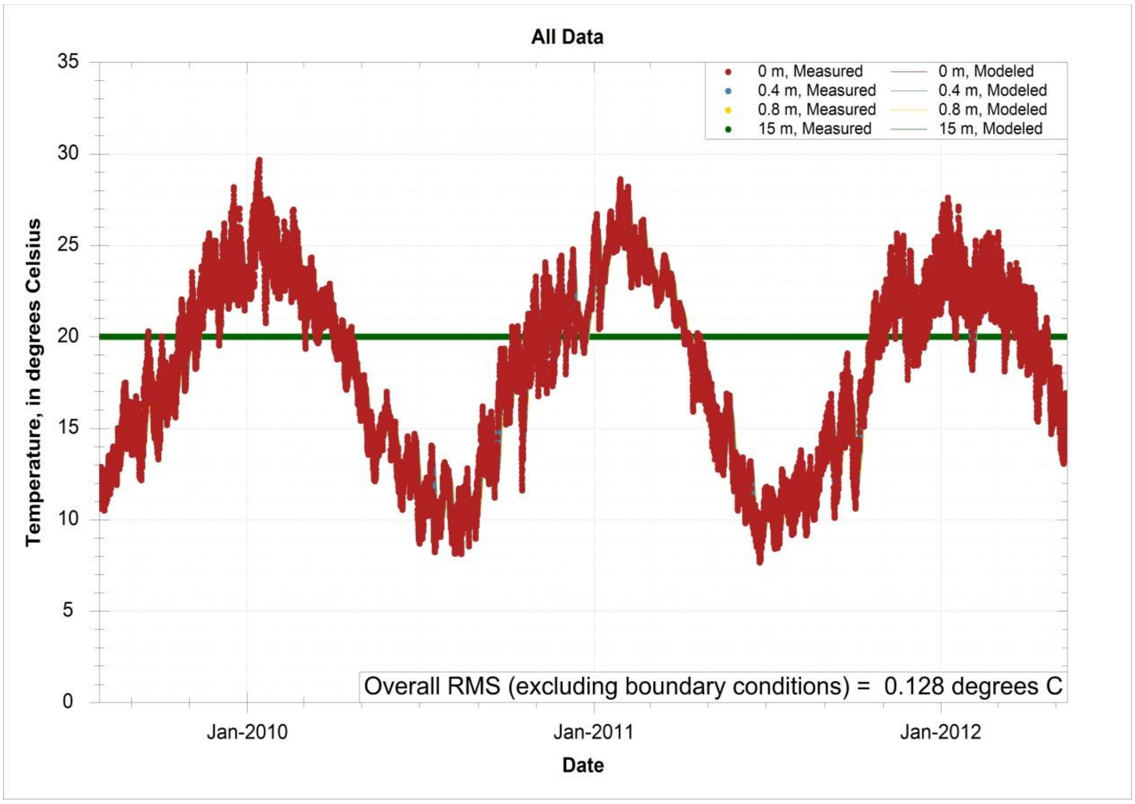


Figure 5-11. Observed and simulated streambed temperatures at different depths.

The overall RMS (excluding boundary conditions) for streambed temperatures measured at 0, 0.4, 0.8 and 15 m depths is 0.128, while the individual RMS values for 0, 0.4, 0.8 and 15 m depths are 0, 0.18, 0.038 and 0 respectively. An RMS value of 0 suggests that a fit is useful for prediction purposes.

5.4.1.1.4 Variability of fluxes at 80cm depth for the entire period (Second Stage)

The second stage of thermal analysis involved the estimation of near streambed fluxes, based on the average hydraulic conductivity calculated in the first stage. The third stage of this exercise explores the dynamic nature of streambed hydraulic conductivity for the entire period, based on streambed bed measurements at 80 cm depth (Figure 5-12).

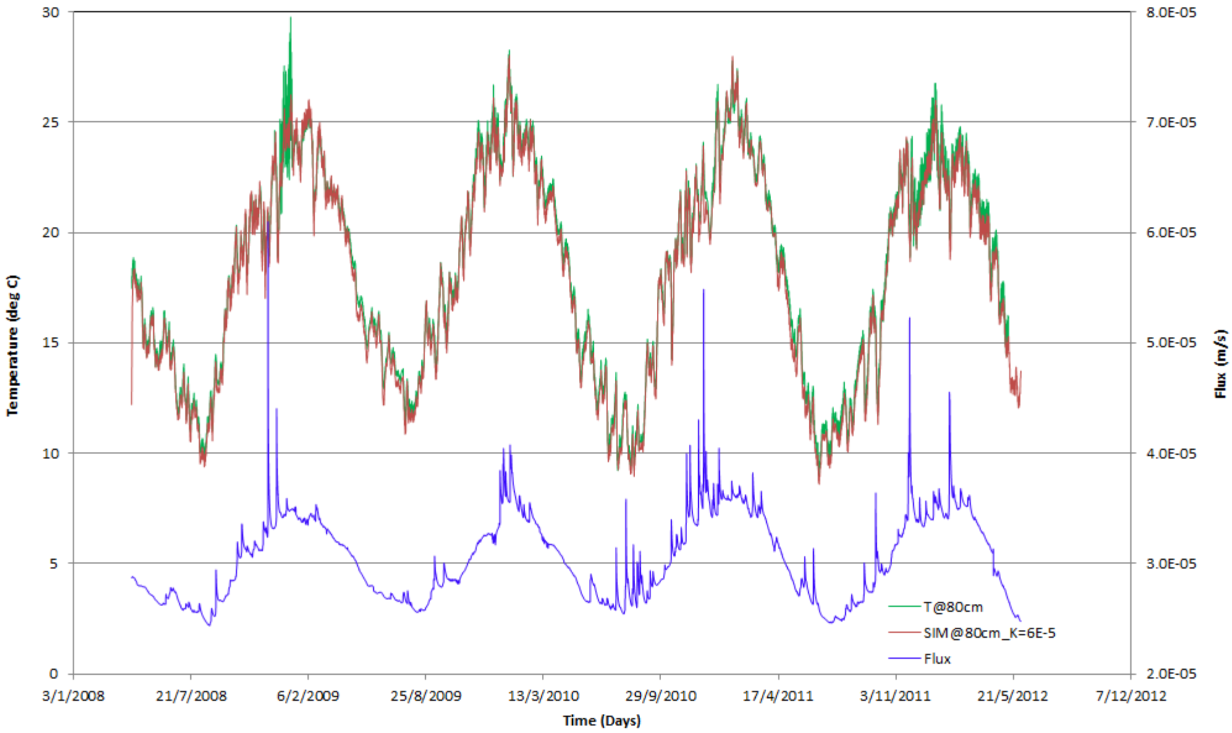


Figure 5-12. Observed VS simulated temperature trends at 80 cm depth and simulated fluxes: Kootingal Gauging Station.

During the observation period, percolation varied from $2.4{\times}10^{-5}$ to $1.3{\times}10^{-2}$ m/s (Figure 5-12).

5.4.1.1.5 Dry spell, prior to 2008 flood events (Third Stage)

A long dry spell prevailed in the study area between 2004 and 2008. The drought was broken by series of stream flow events of different magnitudes and duration that took place in 2008, including two consecutive flood events of November and December 2008. Stream stage varied between 0.57-0.7 meters for the pre-flood events. The near streambed water fluxes ranged from 5.85×10-5 to 6×10-5 (Figure 5-13).

5.4.1.1.6 Period covering the two consecutive flood events in 2008 (Third Stage)

In order to illustrate the dynamic nature of streambed hydraulic conductivity, a more detailed analysis was carried out for the segment representing two consecutive flood events, extending from 19/11/08 to 3/02/2009.

This particular segment represents an extreme flood event that occurred after heavy rainfall over the Cockburn Valley in November and December 2008. These events have resulted in scouring up to 1 m depth upstream of the Kootingal Bridge. As a result of first flood event, two of the three in-stream piezometers, located on the left hand side and middle part of the transect were washed away. However, the left hand side piezometer survived the flood events and provided extremely valuable data set to explore the dynamic nature of streambed conductance at a reach scale.

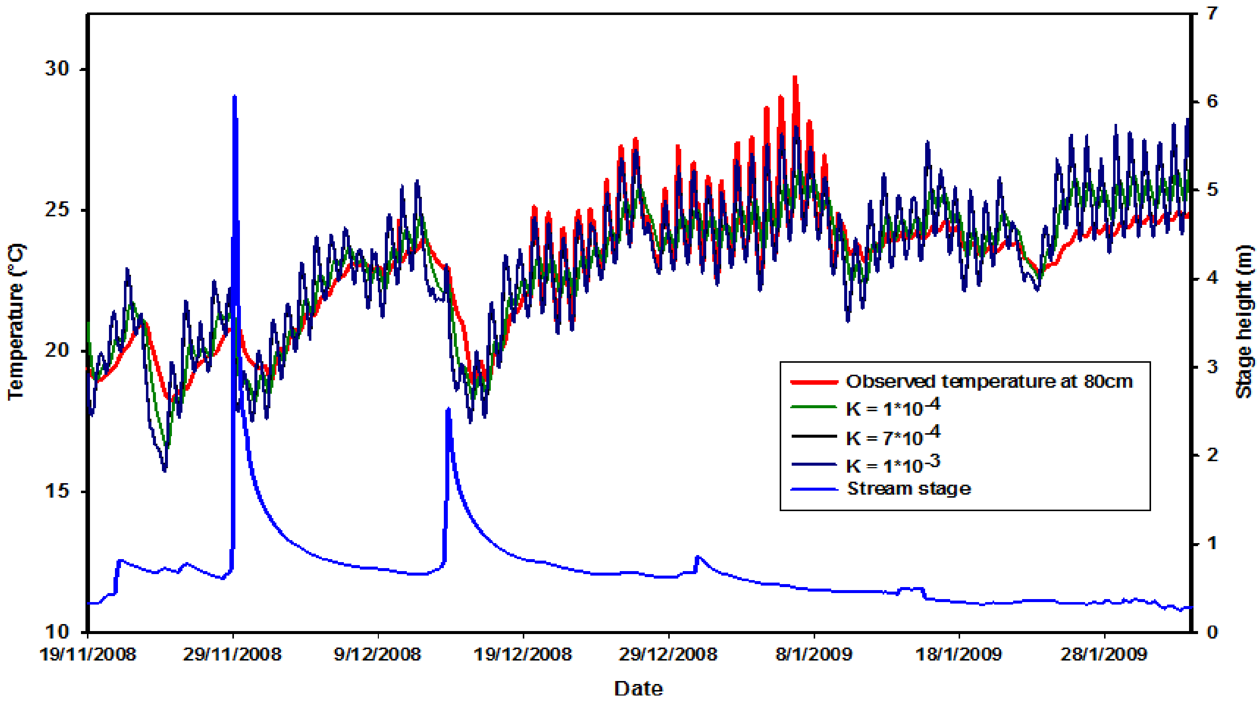


Figure 5-13. Dynamic nature of streambed hydraulic conductivity estimated from heat transport modelling

For simplification purposes, the clogging layer was inserted as an upper layer with a constant thickness of 0.25cm, and the underlying second layer is represents the unsaturated and saturated zone with a maximum thickness of 14.5 m (Figure 4-6). For example, for a 2 layer model with thicknesses of 0.25 and 9.75 meters, a model simulated a composite VHC of 3.5×10^{-6} m/s, generates a VHC of 10^{-7} m/s for an upper clogging layer.

Initially, the VHC values were adjusted by trial and error to minimise differences between simulated and measured streambed temperatures over selected segment of the thermograph. None of the hydraulic conductivities provided a good fit over the entire period of the third segment, extending from 19/11/08 to 3/02/2009. Different sub-segments of the thermograph were fit with different values of VHC, which varied from 8×10^{-5} to 10^{-3} m/s respectively. The best fit value of VHC in the middle part of the segment was 10^{-3} m/s. (Figure 5-10).

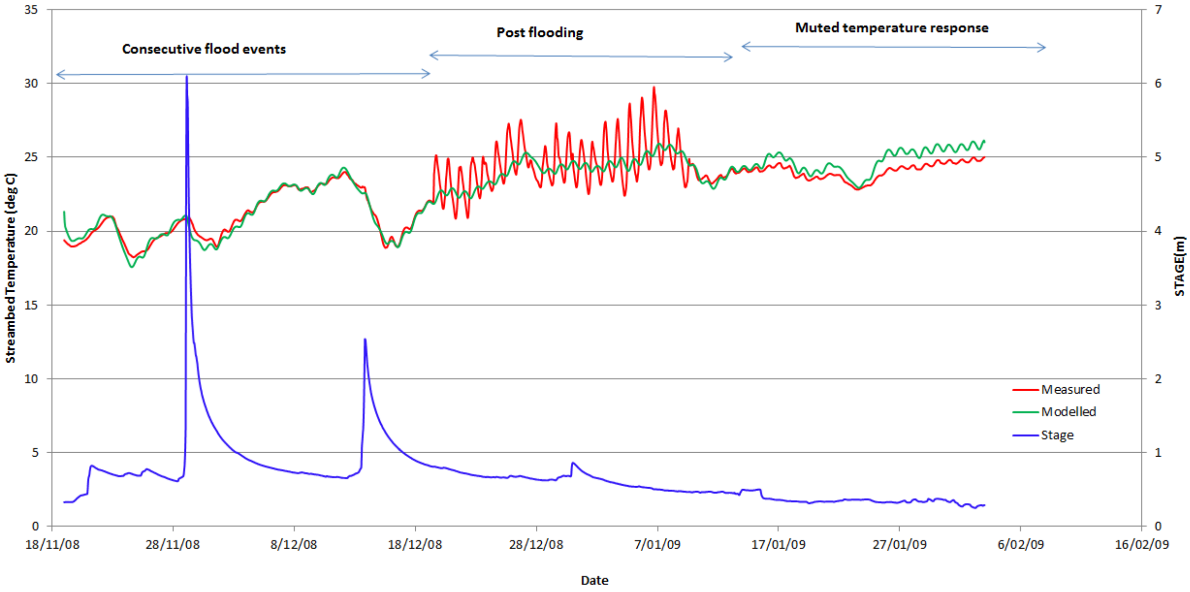


Figure 5-14. Comparison between observed and simulated streambed temperatures at depth of 80 cm below channel elevation using PEST

After attempting to get a best fit between measured and simulated streambed temperatures by trial and error method, an automatic calibration procedure using PEST was conducted for the entire period and the three sub-segments shown in Figure 5-11. As shown in the figure above, the fit between observed and simulated streambed temperatures was reasonably good for the sub-segment representing a pre-flood event. However, the second and third sub-segments showed some discrepancy between measured and observed temperatures, probably caused by the changes in environments. Nevertheless, based on PEST, the vertical hydraulic conductivity for the entire period is was calculated at $4.3{\times}10^{-5}$ m/s. Due to the dynamic nature of the upper streambed layer, the calculated VHC is probably not a true representative value. Thus, each the sub-segments shown in Figure 5-14 were analysed separately in the following paragraphs.

5.4.1.1.7 Sub-segment: Period covering the two flood events

The sub-segment representing flooding event extended from 19/11/2008 to 18/12/2008. During this period the stream stages varied from 0.32 to 6.1 m, while stream temperatures at depth 80 cm below channel elevation varied from 18.2 to 24 °C. For this particular sub-segment the best fit between measured and simulated streambed temperatures was achieved using K_z of 3.9×10^{-3} m/s.

5.4.1.1.8 Sub-segment: Post flooding event

This period extended from 19/12/2008 to 9/01/2009. During this dynamic period, stream stage varied in a narrow range from 0.46 to 0.86m, while stream temperatures oscillated between 20.8 and 29.8 °C. In comparison to the first and third periods, streambed temperature showed high frequency oscillations, probably caused by an increase in *Kz*, induced by break-up of a clogging layer. This process may have induced exchange of water and energy between the stream and the streambed. The simulated K_z for this period was at 4.3×10^{-2} m/s, suggesting an increase in K_z in one order of magnitude, when compared to pre flood event.

5.4.1.1.9 Sub-segment; Muted temperature response period

This period extended from 10/01/2009 to 03/02/2009. During this dynamic period, stream stage varied in a narrow range from 0.25 to 0.50 m, while stream temperatures oscillated between 20.9 and 28.4 °C.

In the third sub-segment, the pattern of streambed oscillations is different when compared with the first two sub-segments. The temperature oscillation was dampened either due to flow of cooler sub-surface (underflow) or relatively cooler water from the RHS bank. The simulated K_z for the third segment was at 3.9×10^{-3} m/s.

Similar reduction in hydraulic conductivity for post flood event was reported by previous investigators

5.4.1.2 Ballantines transect: An Armoured Zone (Pool/Riffle Junction)

5.4.1.2.1 Groundwater level and temperature trends

At the Ballantines transect, the valley floor is about 2 km wide and forms the upper limit of the alluvial aquifer system. The in-channel environment consists of well-developed riffle-pool sequences. However, immediately downstream of Ballantines Bridge the river channel is very shallow and the gravel sediment has been completely removed, exposing bedrock material. There are signs of active channel erosion at this site that are associated with sand and gravel extraction (Galay, 1983).

In order to ascertain magnitude and direction of fluxes, the streambed elevations at the Ballantines transect were surveyed in detail. In the survey, GW93559, located on the left hand side of the bank and two privately owned wells on the right hand side of the bank were included. The available data suggest that the groundwater is far below the channel level for GW93559, indicating a potentially losing stream condition (Figure 5-10).

In the armoured zone near the Ballantines, we have used VS2DHI to estimate the VHC and fluxes. During the two-consecutive flood events of 2009, the VHC varied from 8×10^{-5} to 2×10^{-3} m/s, with stream leakage ranged from 9×10^{-5} to 2×10^{-4} m/s.

5.4.2 Discussion

5.4.2.1 Thermograph trends as a surrogate indicator of a dry spell

In ephemeral and intermittent streams, quantification of exchange of water between the stream and groundwater is problematic. In most of the cases the creek beds are dry and disconnected from the water-bearing formations. In the absence of stream flow data, temperature can be used as a surrogate indicator of dryness in a channel in a qualitative fashion. This approach of inference of stream flow based on pattern of streambed temperature oscillations is very handy in remote parts of Australia, where the density of gauging station is sparse.

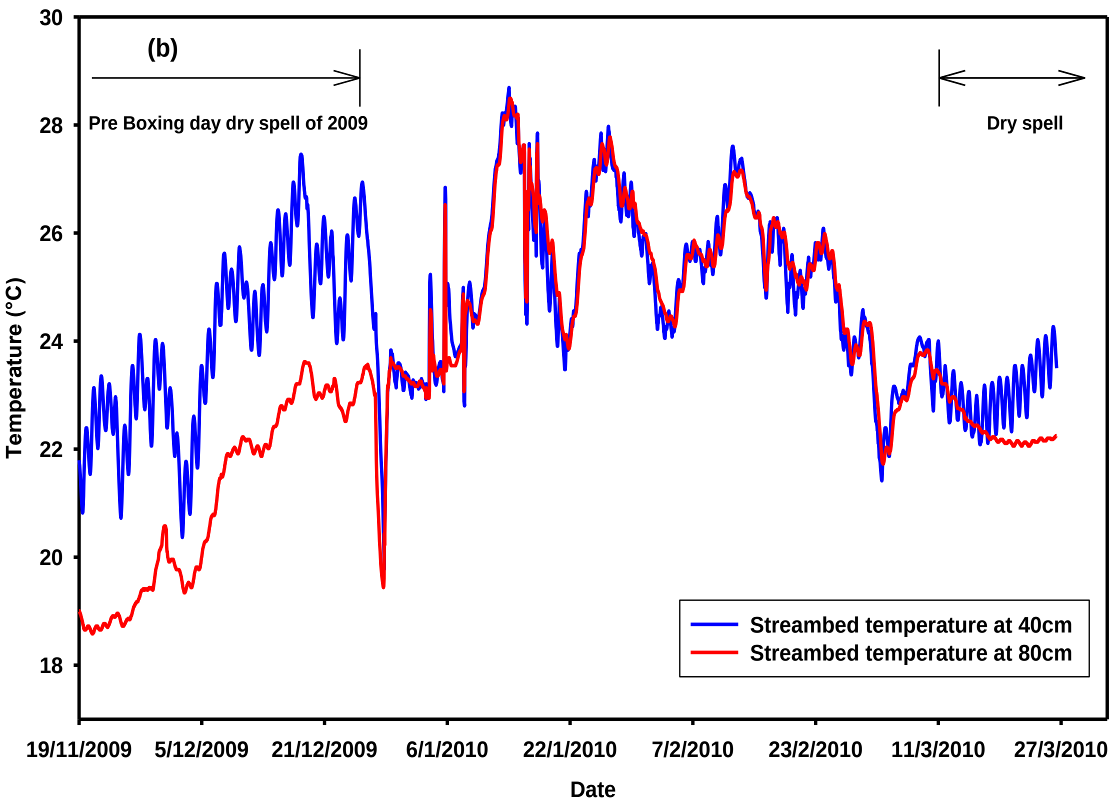


Figure 5-15. Beginning and end of a dry spell, inferred by comparing temperature measurements at two depths at the Ballantines transect.

Based on streambed thermograph trends at two levels (40 and 80 cm below the channel elevation), the degree of connection between surface water and the phreatic groundwater system can be inferred. For example, during runoff events, streambed temperatures at the two measurement points, oscillated in a synchronized fashion, suggesting that both advective and conductive heat transport mechanisms occur. In contrast, during a dry spell, the thermographs at the two levels show different patterns of oscillations (the Cockburn River ceased to flow in November and Dec 2009; the dry spell that persisted in the valley was broken on 27/12/09 by a rare/episodic cyclonic event. Therefore, based on the thermographs patterns at two levels, one could infer the onset, duration and termination of a dry spell for a highly connected, intermittent stream as the Cockburn River (Figure 5-15).

5.4.2.2 Effects of Stream Velocity on Hydraulic Conductivity

There are several factors that affect the magnitude of the streambed conductance. These are: physical, chemical and biological parameters. The physical factors include stream water temperature and velocity of stream flow.

Lyons (1996) calculated average tractive stress duration curves for different bedforms in the Cockburn Valley. In this particular study, it was reported that armoured riffles have a critical stress of approximately 34 N/m2; the probability of occurrence is low and exceeded only 2% of the time during the period of record. Only during extreme hydrologic events would the armoured riffle material be transported as bed load. This confirms our observations that occurred during the last two consecutive flood events of 2008/09, that the in-stream piezometers located on an armoured transect, survived the effect of flooding, erosion and scouring. In contrast, in the pool dominated reaches, the critical tractive stress is approximately 7 N/m2 this level of stress was exceeded 26% of the time. Therefore, the bed material in the pool sites of the study area can be expected to be transported as active bedload 26% of the time. As a result, the in-stream piezometers installed in this section were more often damaged.

During walk over survey, extensive bank erosion was observed in the Cockburn River. Bank erosion is probably associated with bed lowering and changes in land use patterns. In addition, many other factors can be attributed to the cause of bank erosion than bed lowering and natural meandering processes (Lyons, 1996).

5.4.2.3 Effects of Gravel Mining of the Past on Hydraulic Regime and Fluxes

In the context of the Cockburn Valley, an important data set for surface and groundwater connectivity study is the unstable nature of streambed elevation in the valley. Elevation surveys conducted at bridges along the river have shown that streambed of the river has been lowered 0.9 m at Ballantines Bridge since 1965, and up to 2.4 m at Kootingal Bridge since 1939. In order to arrest erosion and streambed lowering, seven weirs (rock ramps) have been constructed on the lower Cockburn Valley (Lyons, 1996).

Despite the contradictory postulates of the past, the effects of bed lowering on water exchange between surface and groundwater are still unclear. However, two scenarios are postulated. For gaining and groundwater-dominated reaches, the lowering of riverbed elevation is expected to increase 'baseflow' and underflow, based on a simple Darcy equation. In contrast, for losing reaches, which are dominant reaches in the Cockburn Valley, stream leakage values generally reach a constant rate when the water table depth is greater than twice the stream width because flow is generally controlled by gravity at these depths (Bouwer and Maddock, 1997).

5.5 Conclusions

The thermal method is a robust and affordable method for estimating infiltration/discharge;

Based on this study, the thermal data indicated that a low flow hydrologic regime creates a favourable environment for clogging of the coarse streambed sediments and the breakup of the sediments takes place during high and flood events;

Streambed thermographs can be used to detect presence of stream flow and this particular application may be useful for environmental flow management;

In conjunction with other geotechnical data sets, an array of streambed measurements in depths can be used to detect of scouring near bridges and natural river systems.

5.6 Recommendations

The recommendations of this charpter are:

- Due to gravel mining of the past, the Cockburn River is still in disequilibrium geomorphological state. Therefore, human induced impacts of the past should be taken into account in any surface groundwater connectivity study;
- Local effects differed from reach behaviour in both space and time. Coupled with:
    - The complexity of geomorphological and geological units,
    - Scale dependence of parameters used in the models,
    - Incomplete knowledge of the hydrological systems
- Upscaling from point scale as in this study to sub-catchment and catchment scale is a challenge that needs to be investigated (Chapter 6) of this thesis.

Chapter 6 Is issue of scale a puzzle in surface-groundwater connectivity investigations?

"Science is often driven forward by the emergence of new measurements. Whenever one makes observations at a scale, precision, or frequency that was previously unattainable, one is almost guaranteed to learn something new and interesting"

Kirchner et al., 2004

Abstract

For water allocation and planning purposes, information on water balance components is required at catchment and sub-catchment scales. However, due to constraints in resources, time and incomplete knowledge of hydro-ecological systems, most of the hydro-ecological investigations are implemented at reach or streambed scales. Thus, synthesising hydrological information obtained at different spatio-temporal scales is a challenge.

As part of a surface water groundwater connectivity study in the Cockburn Valley, I have used heat as an environmental tracer to get insight into hydrological processes that take place at different time/spatial scales. Temperature measured in both in-channel and off-channel piezometers have been used to better understand surface and groundwater connectivity issues at the study site. Within the scope of this study, three 'distinct' spatial/time scales have been conceptualised: Sub-catchment, reach and streambed scale. Based on the available hydro-thermal data sets, the Cockburn Valley is a losing system, at a sub-catchment scale. In contrast, at a finer scale (reach and streambed scales), upwelling and downwelling occurs, mainly induced by local topographic variations. Apart for a small scale reach near the Kootingal Bridge, the extent of interactions between small scale streambed flow paths and the

Kootingal Bridge, the extent of interactions between small scale streambed flow paths and the local groundwater flow system has not been assessed yet. In addition, due to the lack of data, the role of fractured geology on groundwater flow paths is outside the scope of this study. Monthly stress periods are usually used to mimic hydrological process at catchment and sub-catchment scales in humid environments. However, for estimation of recharge or infiltration rates of ephemeral and intermittent river systems in the Namoi Valley, a monthly stress period may not necessarily capture episodic runoff events. Consequently, the effect of temporal domain discretization on infiltration rates was evaluated using an hourly and daily time steps for the Kootingal transect. Within the streambed zone, above the critical water table depth (see Chapter 3), vertical fluxes simulated using an hourly stress period were greater, compared to a daily time discretization.

6.1 Introduction

At a catchment scale, natural groundwater fluxes are controlled by catchment geometry, water table configuration and hydraulic conductivity. Flow system analysis is primarily based on the concept of hierarchical groundwater flow systems (GWFS), for which Toth (1963) first introduced the concept. He was able to show that close relations exist between topography, land use, groundwater recharge, vegetation and soil patterns, groundwater flow paths and surface water networks (Engelen and Kloosterman, 1996; Toth, 2009a) The concept of flow systems is extremely useful in the analysis of spatial and temporal scales of GWFS and their mutual relationships. The subtleties of the influence of topographic relief and climate on GWFS have become better understood in the last 2 decades. Topography or gravity-driven groundwater flow is induced by elevation differences in the water table (1963; 2009b). Flow occurs because hydraulic head decreases from a high-elevation recharge area (high hydraulic head) to a low-elevation discharge area (low hydraulic head). The type of groundwater flow system depends highly on the distribution of topography, the characteristics of the aquifer such as permeability, heterogeneity and anisotropy, and recharge. In the case of arid and semi-arid environments of the MDB, there is a discrepancy between theory and what is observed in the field regarding groundwater flow patterns.

In his conceptual model, Toth (1963) assumed that the water table is a subdued replica of topography. Based on this assumption, three types of flow systems may occur: local, intermediate, and regional. However, a review of hydrogeologic cross-section in different catchments of the northern MDB does not support the occurrence of a nested groundwater flow system. In contrary, water table and the surface topography appear poorly correlated (Desbarats et al., 2002; Haitjema and Mitchell-Brunker, 2005; Moore and Thompson, 1996). Interestingly, regional groundwater flow patterns in the MDB catchments agree well with groundwater simulations, carried out to assess the impact of climate change on groundwater flow regime (Goderniaux et al., 2013). In this particular study, recharge is imposed rather than induced from surface levels.

Taking the issue of scale into context, several investigators (Brunke and Gonser, 1997; Jolly et al., 2008; Tonina and Buffington, 2007; Woessner, 2000) provide insights into small scale hydrological processes that occur at different spatial scales. These processes have become an important hydro-ecological issue over the last two decades as it is obvious that near streambed flows occur simultaneously in vertical and horizontal directions on various scales (Brunke and Gonser, 1997), but a basic unit of exchange occurs across riffle-pool sequences (Berhane et al., 2011; Brunke and Gonser, 1997; Vaux, 1968). On this scale, surface water enters the subsurface sediments at the upstream end of a riffle (a shallow, fast flowing section of a stream) and returns to the stream channel at the downstream end (Brunke and Gonser, 1997; Valett et al., 1994). Interestingly, nested flow systems at a finer, streambed scale are similar to the Tothian GWFS concept highlighted above for humid climatic environments. It is considered to be driven by topographic irregularities at a streambed scale (Elliot and Brooks, 1997; Stonedahl et al., 2010).

Existing surface and groundwater monitoring networks in the Namoi Valley have been revamped during the last decade and this has provided an opportunity to get better understanding groundwater riparian systems. In the context of this study, we have used heat as an environmental tracer to get insight into hydrological processes that take place at different spatial and time scales. Temperature measured in streams, in-channel and off-channel piezometers have been used to understand surface and groundwater connectivity at different spatial and time scales. Consequently, the main objectives of this chapter are to: (1) use temperature profiles beneath streams and in off channel bores to estimate water fluxes using simple analytic techniques and a numerical model; (2) based on field studies, quantify fluxes over a riffle/pool junction at the Ballantines transect; and (3) elucidate surface groundwater connectivity across spatial and time scales (streambed, reach, sub-catchment).

6.2 Study area

The Cockburn River joins the Peel River southeast of Tamworth. In the catchment, the alluvial aquifer of the valley is restricted to a narrow zone between Nemingha and Ballantines Bridge, a length of approximately 15 km varying between 1-3 km in width, and a gradient of 40 m in 15 km. Generally the top 6 metres of the alluvium consist of finer sediments and the main water bearing zones occur below this depth (Calaitzis, 1994). There are a large number of shallow wells and bores in the alluvium, which are generally used for irrigation. The fractured rock aquifer systems are represented by Carboniferous meta-sediments and granites of the New England Batolith. Only few number of bores tap from the fractured rock system. The average annual rainfall decreases with decreasing altitude from the highest point in the Great Dividing Range (1440 m) to Tamworth (280 m). Rainfall ranges from 650 to 800 mm. Tamworth has an annual rainfall of about 670 mm.

The general catchment setting is similar to Chapter 4. In the context of this Chapter, the influences of small scale geomorphological features on temperature profiles and near streambed fluxes was studied in the area between the Mulla Mulla and the Kootingal gauging stations in the upper Cockburn Valley. Small scale processes were captured near the Kootingal Bridge site.

6.3 Review of the physical Processes

Streambed water exchange in the context of this chapter is the movement of stream water from the channel boundaries into the subsurface sediments and back to the channel (Wondzell and Gooseff, 2012). In connected river systems (Chapter 3), stream water may mix with groundwater and the level of mixing in the ‘hyporheic’ zone could range from 0 to 100 %. The physical demarcation of the so called ‘hyporheic’ zone is arbitrary, usually defined by proportions of stream water and groundwater. If the composition of surface water is less than 10% the water is considered to be of groundwater origin (Triska et al., 1989).

Water exchange at a bedform scale is mainly driven by the pressure difference over ranging scales, including local head gradients induced by the interaction of water column currents with sediment bed topography, such as ripples and dunes (Cardenas et al., 2004; Cardenas and Zlotnik, 2003; Elliot and Brooks, 1997a; Thibodeaux and Boyle, 1987). The other form of exchange is induced by propagation of structures (bars, dunes, and ripples). This type of exchange is called bedload movement or turn over exchange (Grant and Marusic, 2011). As bed sediment is scoured from the upstream side of a bedform structure, pore water that was previously trapped in the bedform becomes mixed with stream water (Grant and Marusic, 2011). The dynamic nature of the streambed conductance due to clogging and declogging processes was reported in the previous chapter.

Based on the classification provided by (Kaser et al., 2009), near streambed water exchanges are described by one of the following physical processes (Figure 5.1): (1) Unsteady exchange-between stream bank and channel due to a sudden increases in stream stage (i.e., bank storage processes due to changes in hydrostatic head gradients between stream and lateral riparian aquifer (Lewandowski et al., 2009; Sawyer et al., 2009a); (2) Changes in pressure induced by detachment and reattachment of the turbulent boundary layer over the roughness elements at the sediment-water interface such as cobbles and sediment bedforms (Elliot and Brooks, 1997a; Tonina and Buffington, 2007); (3) Downstream migration of sediment bedforms (Elliot and Brooks, 1997a); (4) Penetration of turbulent sweep and eject events in the sediment (Fritz and Arntzen, 2007); (5) Variation in sediment permeability.

It is beyond the scope of this thesis to capture all the five processes described above in a field environment. Notably, small scale processes (Figure 6-1) are difficult to capture in a field environment, let alone being modeled using traditional Fickian approaches (Gentine et al., 2012). Nevertheless, hydrological data from the Cockburn Valley suggest that water exchanges are probably dominated by small-scale processes, which take place in both a longitudinal and lateral direction.

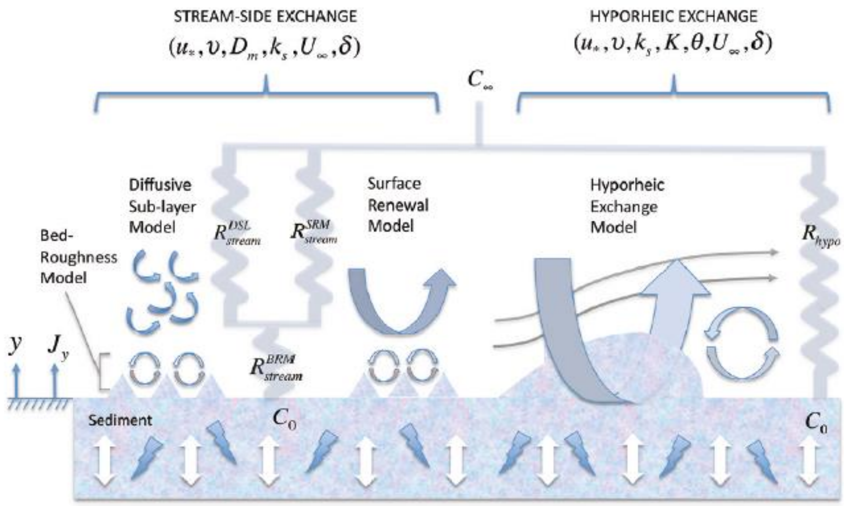


Figure 6-1. Processes thought to control mass transport from the bulk fluid to the surface of the sediment (streamside exchange, left) and across the sediment water interface (hyporheic exchange, right). The arrangement of resistors in the mass transfer circuit is motivated by consideration of the length scales over which different turbulent transport processes operate, as described in the text. The bulk fluid is flowing from left to right. Double sided vertical arrows and lightening bolts represent diffusion (including biodiffusion) and reaction, respectively, in the sediment bed. The illustration for hyporheic exchange depicts detachment of the turbulent boundary layer over a large roughness element, such as a bedform or cobble (After, Grant and Mauric, 2011).

Riffle-scale exchange flows are of particular significance to the stream ecosystem as they are prevalent (Anibas et al., 2012; Wondzell and Gooseff, 2012) and likely account for more interaction (Figure 6-2).

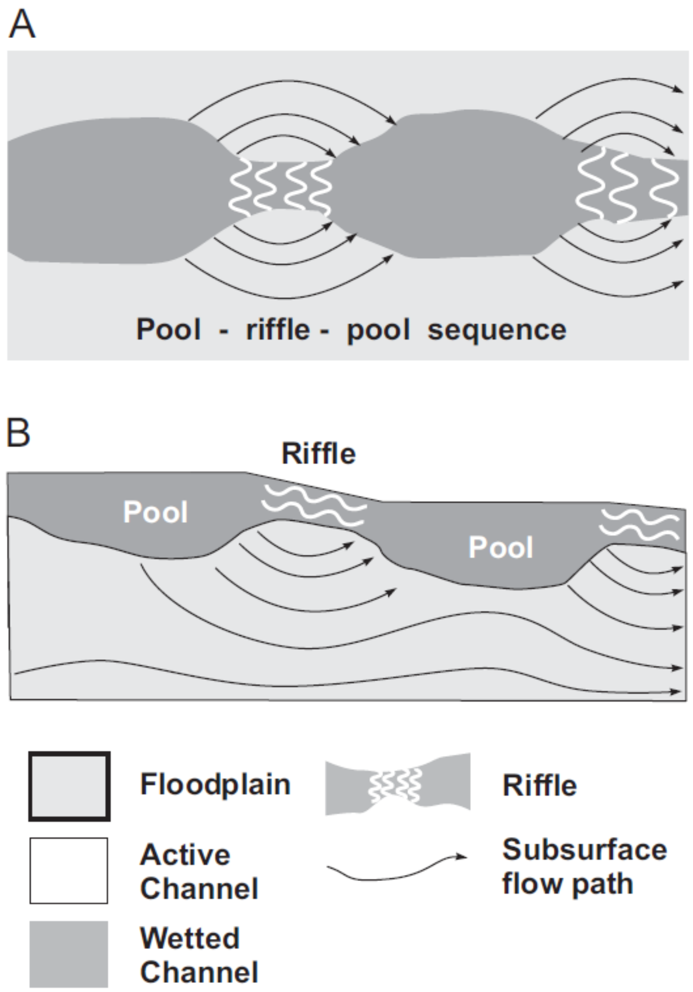


Figure 6-2. Hyporheic flow paths as influenced by pool/riffle sequences. A) Plan view showing hyporheic flow paths through the adjacent floodplains created by the change in the longitudinal gradient over the pool-riffle sequence the amount of HFE is proportional to the head gradient. B) Longitudinal-section along the thalweg of the stream showing the vertical component of HEF flows through the streambed (Wondzell and Gooseff, 1977).

6.4 Methodology

Temperature data was collected at various spatial scales (bedform, reach, and sub-catchment). The instrumentation for both an off-channel and in-channel temperature data collection was easy and straight forward. In the absence of geophysical down the hole logging equipment for this project, available resources were used and improvisation applied when necessary. Six temperature sensors of HOBO® Temperature/Light Data Loggers (at 25 °C, accuracy: ±0.47 °C; resolution: 0.1 °C) type were suspended using a string from the top of bore casings in two monitoring bores GW93038 and GW93039, which are located on the left and right hand side of the bank of the river at the Kootingal transect. The thermistors measured temperatures at an hourly interval in the unsaturated and saturated zones (1, 2.5, 10.5, 14, 14.5 and 15 m below the ground surface), assuming the PVC pipe is in equilibrium with the surrounding environment.

During the first phase of the instrumentation the designated transects in the Cockburn Valley were oriented in a cross-sectional direction. As this setup was not considered an ideal configuration to capture small scale hydrological processes (upwelling and down welling) occurring in a longitudinal direction. Therefore, it was modified in November 2009 by construction of ten additional in-stream piezometers installed in a longitudinal direction along a 20m reach of a riffle-pool transect at the Ballantines (Figure 6-3).

Spot streambed temperature surveys along the Ballantines and Kootingal transects were also carried out using a 'Temperature T bar' (Figure 6-4). It consists of a T shaped handle at the top and a spear point at the bottom, with 5 temperature sensors spaced at distance of 20cm between the spear point and a T bar. The 5 temperature sensors are connected to a DaqPRO 5300 (a 16-bit, high-resolution, eight-channel data logger that offers graphic displays and analysis functions for measuring voltage, current and temperature in real-time). At each site streambed temperatures were measured simultaneously at 5 depths (0, 20, 40, 60, 80, 100cm below channel elevation).

Figure 6-3. Location of in-stream piezometers used to delineate upwelling/downwelling process at the Ballantines transect (picture was taken during a dry spell of 2009).

Figure 6-4. In the same locality as above, a T bar device was used to measure streambed temperatures at different depths. During flow events, it was not easy to find a plastic hose connected to the upper opening end of a buried piezometer pipe, which sticks out about 5 cm above the streambed

At a finer scale, a temperature profiler was used to collect data at a beform scale. A more detailed description of this instrument is provided by (Anibas et al., 2012; Schmidt et al., 2006). Streambed temperature mapping using a temperature profiler is simple and robust in data collection and analysis (Anibas et al., 2012; Bredehoeft and Papadopulos, 1965a).

6.4.1 Analytic method

6.4.1.1 Bredehoeft and Papadopoulos (BP) method

The Ex-Stream package, which integrates coupled heat and fluid flow analytical models (Hatch et al., 2010; Schmidt et al., 2006; Stallman, 1965; Suzuki, 1960) was used for the delineation of reaches into gaining/losing and estimation of fluxes at a streambed scale. Among these methods, the Bredehoeft method (BP) is of interest in the context of this study. It was originally applied for the estimation of fluxes at a basin scale, where a steady state condition is a prerequisite for its application. However, the BP method was considered to be the least data demanding and simplest to apply within the Ex-Stream package. The assumption was that the streambed exchanges were dominated by vertical flow at the scale of the measurements and the BP equation could be applied, as it is based on the one dimension heat transport equation of the form:

$$k\frac{\partial^2 T}{\partial z^2} - v_z \rho_w c_w \frac{\partial T}{\partial z} = \left[n\rho_w c_w + (1-n)\rho_S c_S\right]\frac{\partial T}{\partial t} \qquad \text{Eq. 6-1}$$

Bredehoeft and Papadopoulos (1965) derived an analytical solution of the above equation and the solution is represented by:

$$\frac{T - T_0}{T_L - T_0} = \frac{\exp(\beta z/L - 1)}{\exp(\beta) - 1} \qquad \text{Eq. 6-2}$$

$$\beta = \frac{c_w \rho_w v_z L}{k} \qquad \text{Eq. 6-3}$$

Therefore the fluid velocity v_z can be calculated from:

$$v_z = \frac{k\beta}{c_w \rho_w L} \qquad \text{Eq. 6-4}$$

Where β is the Peclet number; T_z is the temperature at any depth z; T_0 and T_L are the uppermost and lowermost temperature measurements respectively; L is the length of the vertical section over which the temperature measurements are made; v_z is the vertical fluid velocity; c_w is the specific heat of the groundwater; k is thermal conductivity; ρ_w is the density of the groundwater.

Traditionally, in regional groundwater studies, type curves are matched to a temperature profile measured in a well penetrating the confining bed was used to calculate vertical flux. Vertical ground water velocity (v_z) is computed from the Peclet number (Eq. 6-4), which is read from a type curve. Although the BP model was developed for analysing profiles across a confining bed within the geothermal zone, the model is applicable to other one-dimensional problems where temperature is constant at either end of the system (Anderson, 2005).

Recently, the BP method has been used for small scale investigations (Swanson and Cardenas, 2011). For example, in the Ex-Stream package, the Bredehoeft method modules use period averaged temperature profiles to solve explicit equations for vertical fluid flux (Swanson, 2010).

The Peclet number is a dimensionless parameter that compares the proportion of heat transported by advection to heat transported by conduction. Domenico and Palciauskas (1973) suggested that when the Peclet number is substantially less than 1, conductive transport dominates. Conversely, when the Peclet number is substantially greater than 1, then advective heat transport dominates. A concave temperature profile below the depth of seasonal fluctuations of temperature indicates groundwater recharge taking place, with the degree of curvature related to the magnitude of recharge. Conversely, a groundwater dominated system is expected to exhibit a convex profile (Bredehoeft and Papadopulos, 1965b).

6.4.1.2 Stallman's analytical solution

For quick visualisation of monthly temperature profiles under streambed, analytical method developed by Stallman was used. A governing equation for one-dimensional, vertical, anisothermal flow of an incompressible fluid through a homogenous porous medium with a sinusoidal temperature is presented by the following equation (Stallman, 1965; Suzuki, 1960):

$$K_t \frac{\partial^2 T}{\partial z^2} - qC_w \frac{\partial T}{\partial z} = C_m \frac{\partial T}{\partial t} \qquad \text{Eq. 6-5}$$

Where T is temperature in degree Celsius; z is vertical depth below streambed, C_b and C_w are the volumetric heat capacities of the bulk medium and water, K_t the thermal conductivity of the bulk medium and t is time in seconds.

Daily or annual temperature fluctuations can often be approximated by sinusoidal functions (Nimmo et al., 2005). If recharging fluxes are steady and materials homogeneous, the recharge rate *q* can be obtained by inversion from the phase shift (*b*) and attenuation (*a*) of the temperature waves as they propagate downward from the land surface (Stallman, 1965):

$$T(z) = T_0 + \Delta T \exp(a - z) \sin\left(\frac{2\pi}{\tau - bz}\right) \qquad \text{Eq. 6-6}$$

where *a* and *b* are related to thermal-conduction and advection constants *K*' and *V*' by

$$a = \left[\left(\frac{k' + V'^4}{4}\right)^{1/2} + \frac{V'^2}{2}\right]^{1/2} - V' \qquad \text{Eq. 6-7}$$

and

$$b = \left[\left(\frac{k' + V'^4}{4}\right)^{1/2} + \frac{V'^2}{2}\right]^{1/2} \qquad \text{Eq. 6-8}$$

Here T_0 and ΔT are the mean and amplitude of the temperature signal at the land surface, and *V*' and *K*' are defined as

$$K' = \frac{\pi C_b}{k_b \tau} \qquad \text{Eq. 6-9}$$

and

$$V' = \frac{q C_w}{2 k_b} \qquad \text{Eq. 6-10}$$

Stallman's analytical solution depends on measurements of streambed temperatures at two depths for calculating amplitude attenuation and phase lags between the diurnal temperature series.

6.4.1.3 Extinction depth

Despite the fact that air/stream temperature in arid and semi-arid environments exhibit high amplitude diurnal fluctuations, the temperature signal on a daily scale does not propagate deep into dry channel sediments. The damping depth (Z_d) is a characteristic depth, at which the temperature amplitude reduced by e^{-1} (1/2.718=0.37). The damping depth is related to the thermal properties of the soil and the frequency of the temperature fluctuations as follows:

$$Z_d = \sqrt{\frac{2D_T(\theta)}{\omega}} = \sqrt{\frac{2k(\theta)}{\omega C^*_B}} \qquad \text{Eq. 6-11}$$

Where ω is the angular frequency of diurnal temperature oscillation, which equals to 2π/86400. As the effective thermal diffusivity, D_T, can be expressed as the ratio of thermal conductivity to volumetric heat capacity, therefore, Z_d is a function of soil bulk density and water content (Blume et al., 2002) . The decrease of amplitude and increase of phase lag with depth are typical phenomena in the propagation of a heat wave in the unsaturated zone (Hanks and Ashcroft, 1980).

6.4.2 Numerical method

The numerical model VS2DH (Healy and Ronan, 1996a) was used to visualise monthly temperature profiles to delineate gaining and losing reaches in the Cockburn Valley. A more detailed description of the package can be found in numerous USGS publications (Healy and Ronan, 1996b; Hsieh et al., 2000a; Lappala et al., 1987) (see Chapter 3 of this thesis).

6.5 EM survey

In order, to study the textural distribution of the in-stream sediments, an EM31 survey was carried out on the 14th December 2009, when the streambed was relatively dry. We used a Geonics EM31 ground conductivity meter, mounted on a vehicle to measure the apparent conductivity of the in-streambed sediments. The EM instruments measure the ratio of strength of a primary electro-magnetic field and a secondary inductive field. This ratio depends on the electrical conductivity of the bulk soil. They can be operated in two modes, a horizontal mode in which effective penetration depth is the smallest and in vertical mode where penetration depth is greater. For this study, the EM survey was carried out in a vertical mode with a maximum penetration depth of 6m. Advantages of the EM are that measurements are quantitative, can be taken rapidly, do not require physical contact between the instrument and the soil and are less influenced than other techniques by the strong micro scale spatial variability that is characteristic of soil salinity, but can also depend on variations in water content and clay content.

6.6 Results and analysis

6.6.1 EM Survey

EM Survey result is shown in Figure 6-5, where the apparent conductivity varied from 180 to 250 mS/cm. The highest conductivities are assumed to be associated with gravels, cobbles and boulders laden with fine sediments, while ECa measurements in the lower scale are assumed to be associated with fresh water lenses under the channel elevations (shown in blue).

The EM data indicates significant small-scale spatial variability at the stream bed scale. In addition, it can be assumed that the locations with fresh water lenses would have higher stream water exchanges than the other locations.

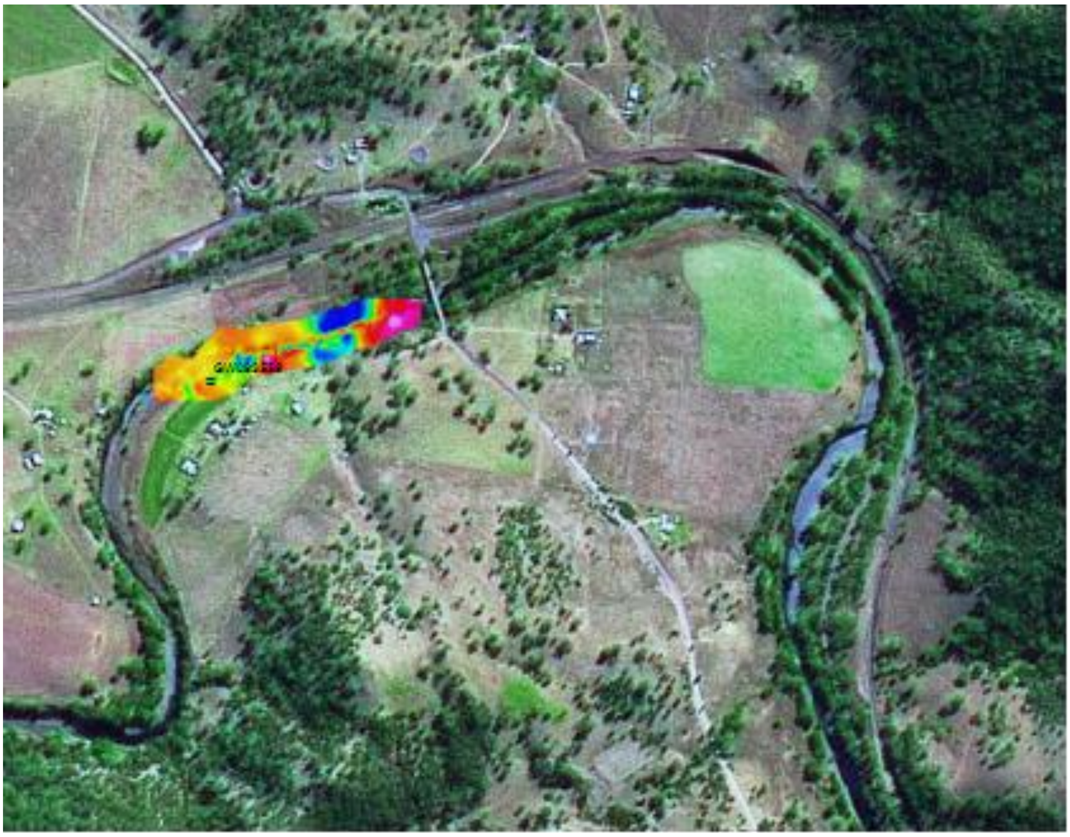


Figure 6-5. An EM-31 map for a dry spell period overlayed over a Spot 5 image for a riffle/pool system at the Ballantines transect.

6.6.2 Bedform scale

The interaction of flow and moveable sediments often creates bedforms such as ripples, dunes, antidunes, and bars, as was also picked up in the EM survey. These bedforms in turn can interact with the flow to modify the rate of sediment transport Allen (1984) suggested that the sequence of bed configurations depends on the Froude number (Fr). For a given alluvial sediment bed, dunes and antidunes are predicted to form only for $Fr^2 < \tanh(kH)/kH$ and $Fr^2 > \tanh(kH)/kH$ respectively, where Fr is the Froude number of the flow, $k = 2\pi/\lambda$ is the wavenumber, λ is the wavelength of the bed waves, and H is the undisturbed mean flow depth (Kennedy, 1969).

Due to hydrological complexity at a finer scale, some investigators prefer to couple surface and groundwater processes (Tonina and Buffington, 2007; Tonina and Buffington, 2011). For simplicity, hyporheic flow is represented by groundwater flow equation of the form: $u = k\nabla h/n$, where *u* is the seepage velocity, *k* is the hydraulic conductivity, *n* is the sediment porosity and h is the piezometric head.

In the vicinity of the Kootingal transect, two types of near streambed water exchanges have been mapped. Groundwater bank 'discharge' probably dominates the right hand side of the bank of the Cockburn River (Chapter 5 of this thesis). In contrast, on the left hand side of the bank, the near streambed water exchanges are more complex and appear to be dominated by small scale features as sand bars, ripples and dunes. These two complex hydrological processes were captured by streambed temperature survey carried out on 14^{th} June 2010. The survey was carried out both in a longitudinal and lateral transects on the same day (Figure 1), when the difference of an ambient groundwater temperatures in contrast to the surface temperature was considered to be extreme.

6.6.3 Lateral transect

Figure 6-6, shows streambed temperature variations along 14 m wide lateral cross-section in a pool system, about 300 m upstream of the Kootingal Bridge. Surface water temperature ranged from 14 to 16.5, while streambed temperature measured at a depth of 1m meter ranged from 15.9 to 19.9^{o} C. Based on qualitative interpretation of the thermal data set, three features are notable. First, the thermograph data display a wave like temperature response, probably linked to the depositional environment of the sediments in the study area.

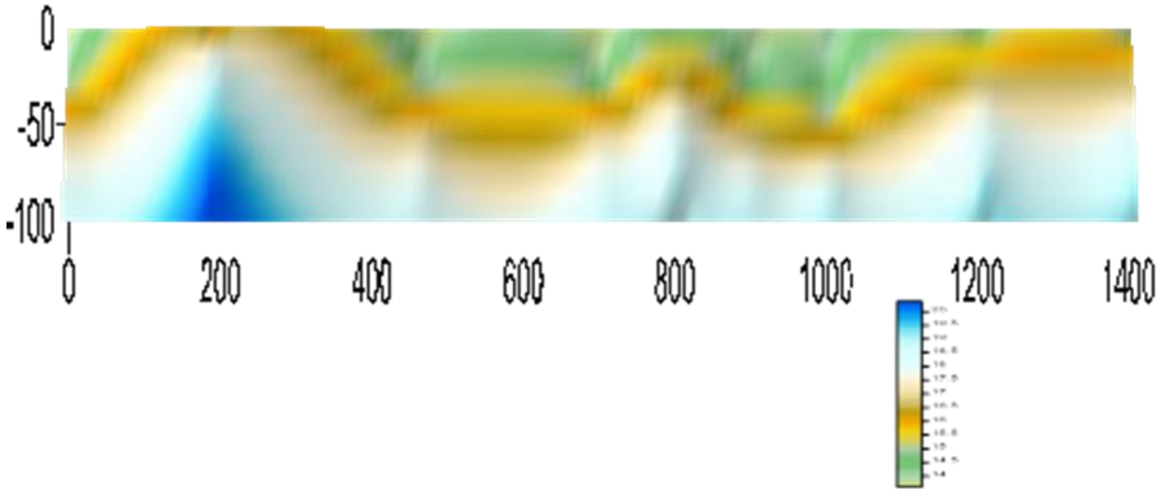


Figure 6-6. Streambed temperature variations along 14m long lateral cross-section, about 500m upstream of the Kootingal Bridge (Vertical and horizontal distances shown are in cm)

Second, an anticline structure, which appears on the left hand side of the cross-section, is discontinuous and probably was eroded during high runoff events or flood events. In this part of the reach, more gravel bed sediments are eroded from the system, due to the gravel mining activities in the past (Berhane et al., 2011). Third, at about 200 cm distance from the right hand side of the bank, a positive temperature anomaly of an upwelling zone was mapped. This localised groundwater discharge zone might be induced by elevated groundwater heads in comparison with stream stage on the right hand side of the bank.

6.6.4 Longitudinal spot survey

Based on comparison of surface and groundwater heads, the Kootingal reach was characterised earlier as a through flow system, gaining on the RHS of the bank and losing on LHS of the bank (Berhane et al., 2011). Thus, due to its hydrological complexity, the Kootingal transect was targeted for detailed spot measurements. Figure 6-7, shows representative locations for the longitudinal spot temperature survey carried out on 14th June

2010. Results of spot streambed temperature survey targeting individual bed forms in a vertical profile are shown Figure 6-8.

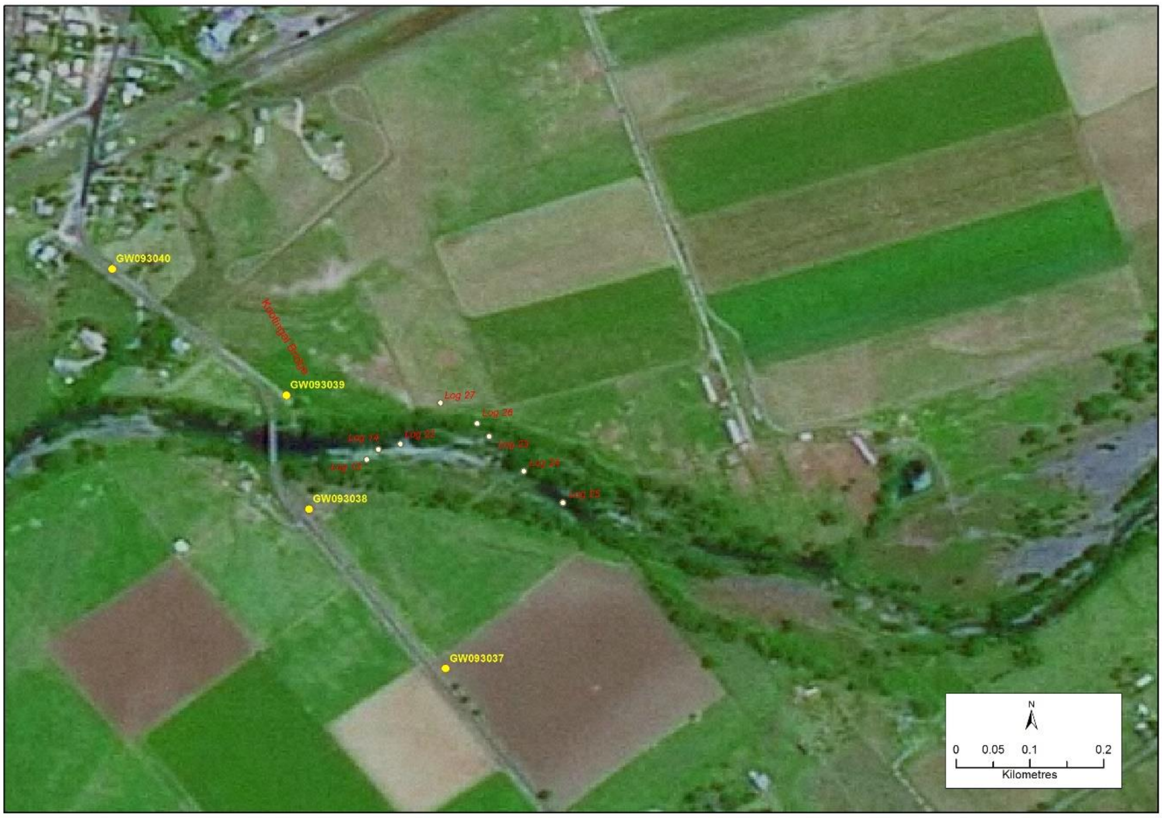


Figure 6-7. Locations of spot streambed temperature measurements, upstream of the Kootingal transect (GW093037-GW03040= piezometer; Log 13= Sandbar (LHS); Log 24= Tail of riffle; Log 25= Pool site; Log 26 and 27= groundwater seepage zones)

Although the measured temperature profiles are within the surficial zone and are highly transient as a function of solar radiation, air temperature, stream temperature and surface and groundwater pumping, the Bredehoeft method was applied for analysis of temperature profiles. Qualitatively, three types of streambed thermal responses within the streambed zone have been identified: (1) A convex profile related to an upwelling zone; (2) A concave profile related to a downwelling zone; (3) A complex profile, mapped in the pool site, probably dominated by mixed temperature signatures.

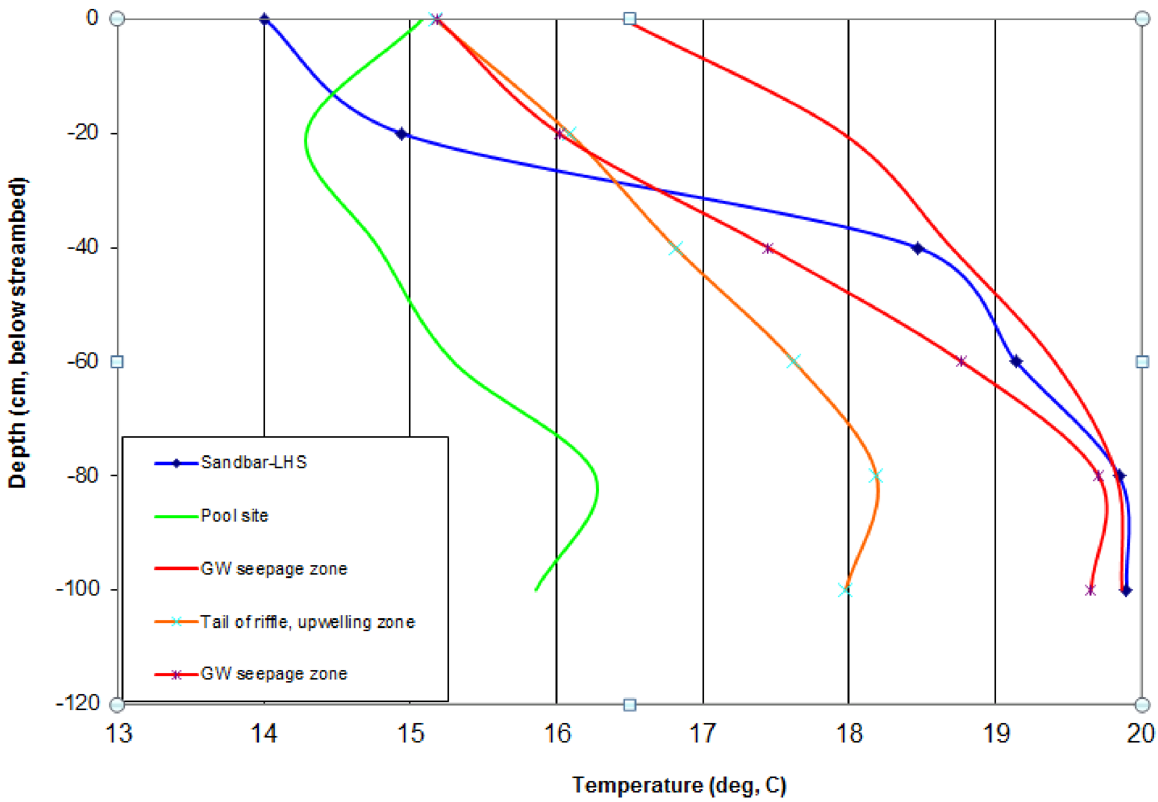


Figure 6-8. Depth temperature profiles in different bedforms: Upstream of the Kootingal transect

Concave type responses were predominantly mapped on the RHS of the bank (Profile 1, 2, 4 and 5), are considered to be influenced by a localised groundwater discharge zone, prevailing on the RHS of the bank. Some of the profiles showed a complex temperature response, due to transient nature of heat and water fluxes in the upper part of the streambed (pool and tail of the riffle site). During this reconnaissance survey, all of the profiles exhibit a convex shape, below 80 cm depth.

After qualitative analysis of temperature profiles, we have used the BP method to calculate instantaneous magnitude of fluxes, using the Ex-stream analytical package. The results of the analysis are given in table below. The calculated upward fluxes varied from -5.28 to -6.98*

10^{-3} m/day on the RHS of the bank representing contrasting bedforms. During the survey, downward flux occurred only at the tail of the riffle site (Table 6-1). Due to transient nature of the system, the calculated fluxes should be considered as point value in time. The magnitude of flux depends on stream stage and stream temperature. Downward fluxes are shown as positive values and upward fluxes are shown as negative values.

Table 6-1. Simulated fluxes for different streambed units

Streambed type	T_o (°C)	T_z (°C)	T_1	Flux m/day (10^{-3})
LHS of the bank (Log13)	14	19.1	19.9	-6.887
Pool site	15.17	17.62	17.97	-6.981
Tail of riffle	15.08	15.29	15.85	1.557
GW seepage	15.18	18.77	19.65	-5.282

T_o = Streambed sensor.

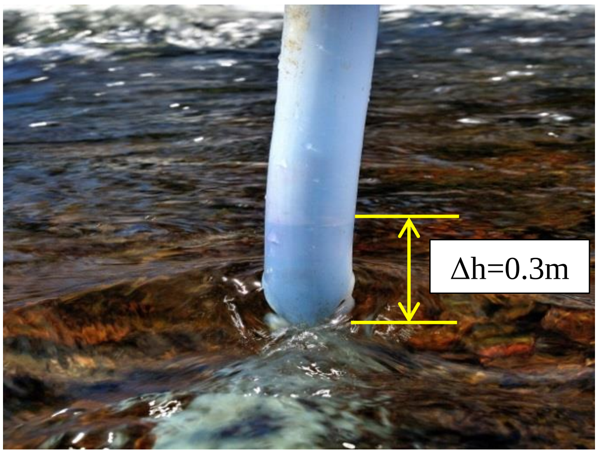
6.6.5 Riffle/pool junction

Although the unravelling and delineation of multi-scale interactions is problematic in this study, near streambed exchanges can occur simultaneously on various scales (Brunke and Gonser, 1997). Channel features such as riffles may establish longitudinal “hyporheic’ flow and temperature patterns as demonstrated by flume experiments (Savant et al., 1987; Thibodeaux and Boyle, 1987; Vaux, 1968)

There are limited case studies in literature on the depth penetration of stream water, which can be inferred from streambed temperature patterns (Hartman and Leahy, 1983; Harvey and Bencala, 1983) at a pool/riffle junction. Most of the information cited in the literature is based on controlled laboratory studies. In contrast, an upwelling component at a riffle junction of the Ballantine transect was captured during the spring runoff event of 2010, when a relatively colder surface water entered the subsurface sediments at the upstream end of a riffle (a shallow, fast flowing section of a stream) and upwelling of relatively warmer streambed water occurred at 600 cm. The observed streambed isotherms were confirmed with Vaux's model (Figure 6-9).

Based on a Darcy's Law ($q_{flux} = -k\frac{\Delta h}{\Delta l}$), where: q_{flux}, is vertical flux, k is the saturated hydraulic conductivity, $\frac{\Delta h}{\Delta l}$ is the vertical hydraulic gradient. The calculated upward flux is in the range of 10^{-7} m/s. Clearly, high-gradient streams with coarse-textured bed sediment (large k) have a great potential for 'hyporheic' exchange. Conversely, low-gradient streams flowing over fine-textured bed sediment have a much smaller potential for 'hyporheic' exchange (Storey et al., 2003).

(a)



(b)

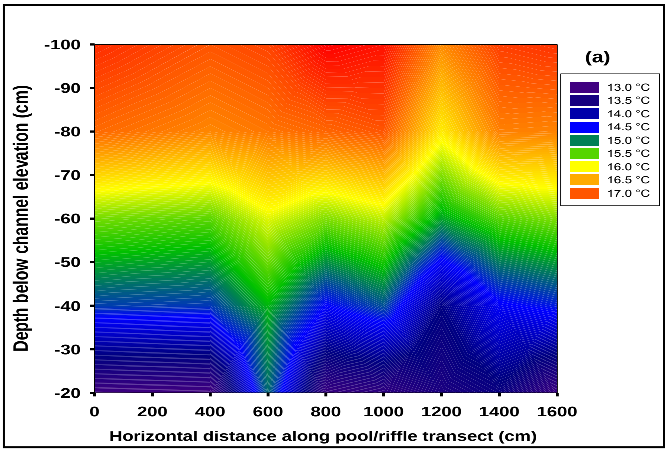


Figure 6-9 . (a) Occurrence of a positive head of 6 cm at a tail of riffle; (b) Streambed isotherms along the Ballantines transect showing upwelling at 600cm and downwelling at 1200cm measured using a Temperature T Bar.

6.6.6 An intermediate spatial scale (reach)

6.6.6.1 Stream temperature and stream stage trends

Due to the occurrence of single and multiple hydrologic events at the Kootingal gauging station, the 2011 water year was selected for detailed analysis and development of bore temperature envelopes (Figure 6-10). As a warming up segment, December 2010 was included in the analysis. For summer and autumn season (1st December 2010 to 31st May 2011), stream stage varied from 0.26 to 2.5 m, while stream temperature varied from 12.3 to 28.2 °C. The propagation of heat wave in depth in the streambed zone is presented in Figure 6-14 to Figure 6-26.

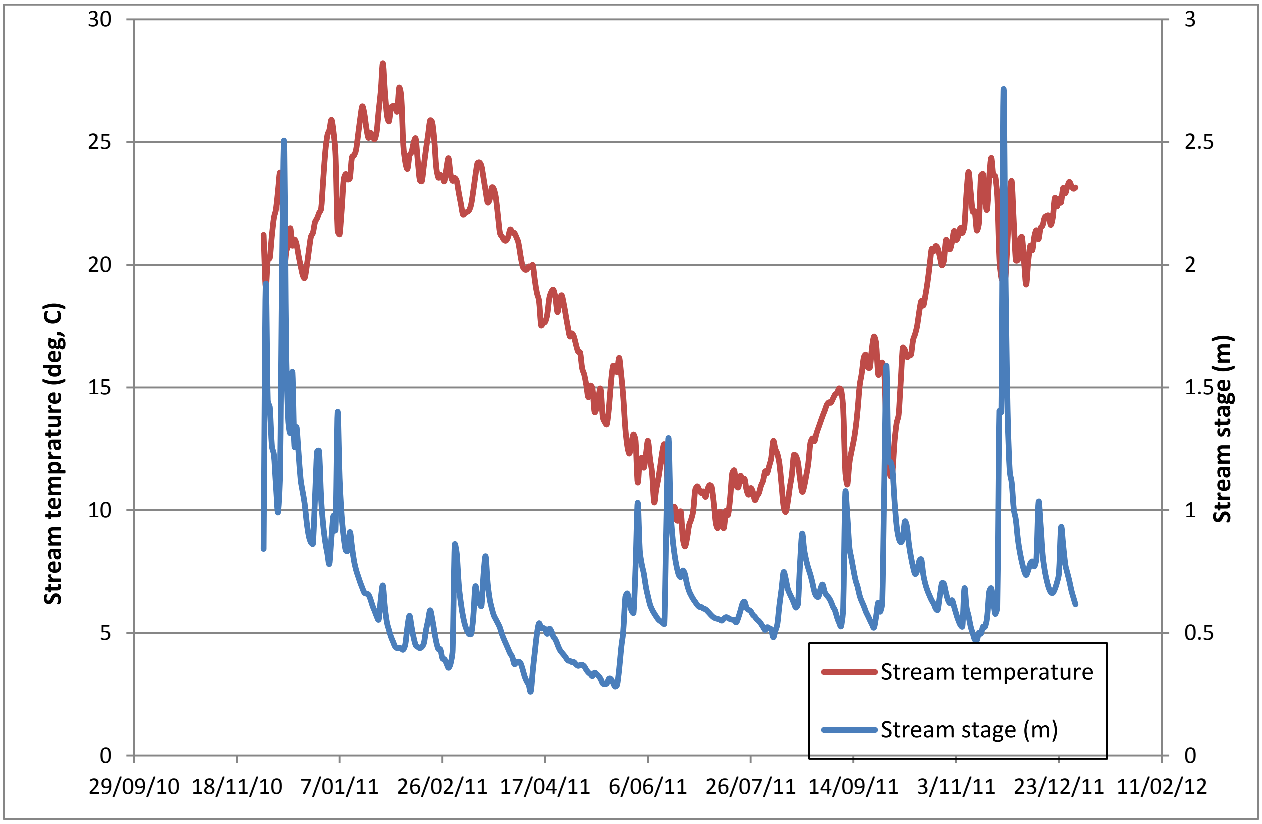


Figure 6-10. Kootingal gauging station hydrograph and thermograph for 2011.

During the second half of the year, representing winter and spring (1st June – 31st November 2011), stream stage oscillated between 0.47 and 2.7 m, while stream temperature varied from 8.5 to 24.3 °C.

6.6.6.2 Development of bore temperature envelopes

In order to get insight into surface groundwater connectivity in space and time at a reach scale, hourly temperature measurements at different depths were carried out in off channel monitoring bores (GW93038 and GW93039, Figure 6-38).

Prior to presenting results of measured bore temperature profiles from the Kootingal transect, theoretical bore temperature profiles were generated for gaining and losing reaches, based on analytical solutions (Fritz and Arntzen, 2007; Lapham, 1989a; Stallman, 1965). The streambed hydrothermic properties used for simulations of the monthly temperature profiles are given in Table 6-2. For both gaining and losing reaches the hydrothermic properties remain the same, the only variable parameter that changes sign is the flux component. In this convention, a negative flux presents a losing reach, while a positive flux infers a gaining reach.

Table 6-2. Hydrothermic properties used for simulation of monthly temperature profiles

Delta T	**Period**	c_w	ρ_w	c_{total}	ρ	χ	q_z (m/s)
5	31536000	4.18E+03	1000	850	220	3	1.15E-07

Description of parameters is provided in section 4.3.

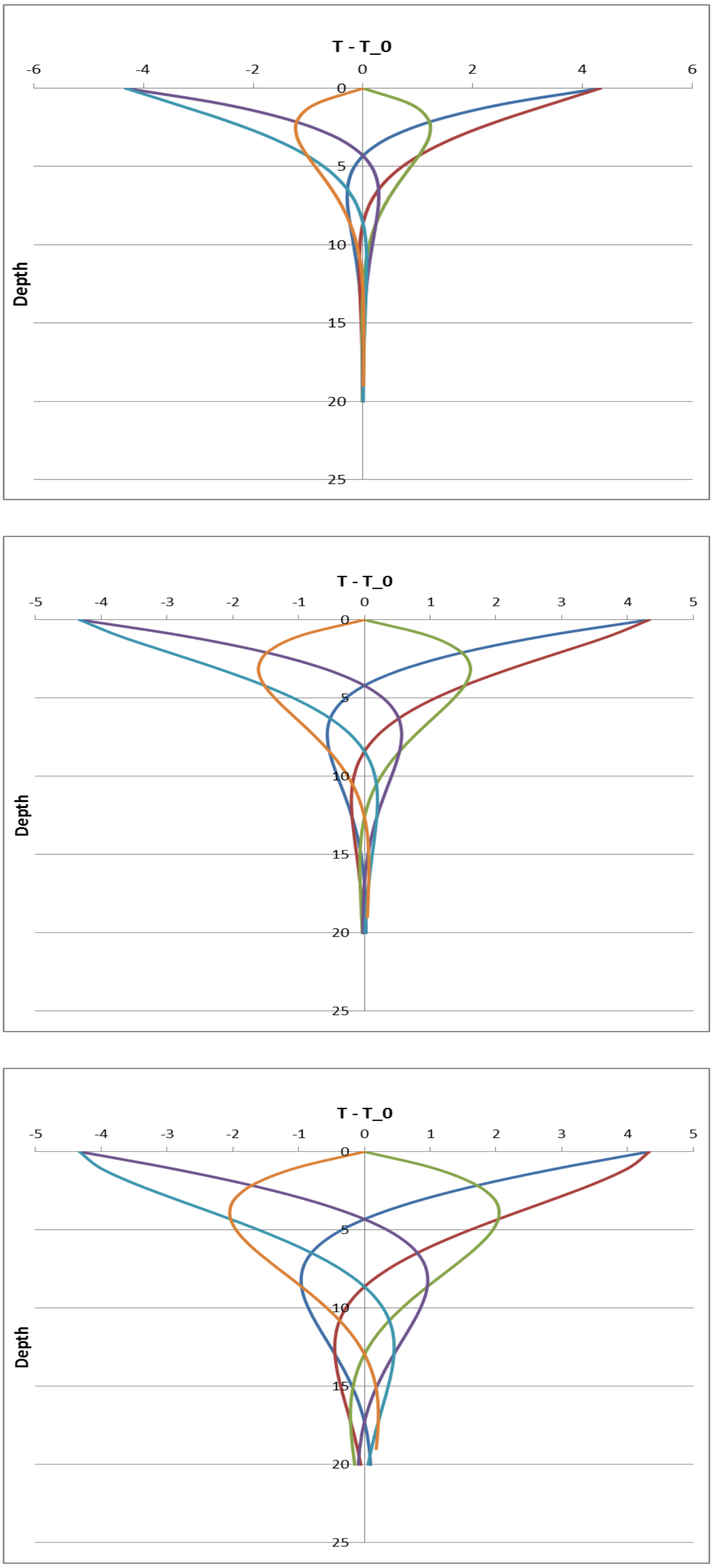


Figure 6-11. Theoretical monthly temperature profiles for losing (a) gaining (b) neutral and (c) losing reaches, depth in the y-axis is in cm.

By inserting appropriate parameters in Eq. 6-11 the damping depth ranges from about 12 cm for dry sediments to 22 cm for wet sediments. However, it should be noted that the equation doesn't take into account advection induced propagation of thermal wave. A closer look at Figure 6-11 shows that heat wave propagates deeper into the profile and monthly temperature profiles stretches downward beneath a losing reach (Figure 6-11a), due to the absence of any moderating effect due to groundwater inflow (Constantz and Stonestrom, 2003a) . In contrast, in a gaining reach, the annual envelope temperature is stretched upward towards the streambed surface (Figure 6-11b). In neutral reaches of a stream, the shape of temperature profiles is mainly controlled by thermal conduction.

Having these three distinct thermal characteristics for gaining, neutral and losing reaches in mind, temperature profiles for the Kootingal transect were analysed. Temperature envelopes measured in GW93038 and GW93039 are shown in Figure 6-12. On both sides of the river bank, heat wave propagation extended up to 16 m below the surface. The reader should note that this is the maximum extent of the unconsolidated sediments in the study area (Appendix A). As expected temperature profiles in GW93038, stretched downwards on the bank adjacent to the losing reach. However, the shape is more complex, consisting of two components, a cone like shape in the upper and unscreened part of the bore casing, a narrow loop like structure in the screened section of the bore. Noticeably, the shape of the temperature profile in GW93039 is similar to GW93038, with slight distortions in the screened part of the profile, likely caused by human interferences associated with placement thermistors at the right depth, after they have been taken for down load. The shape of temperature envelope in GW93039 is determined by the following three factors: (1) direction of groundwater flow (v_x or v_z); (2) horizontal distance of the bore from the channel (retardation of heat) and (3) magnitude of infiltration via the unsaturated zone.

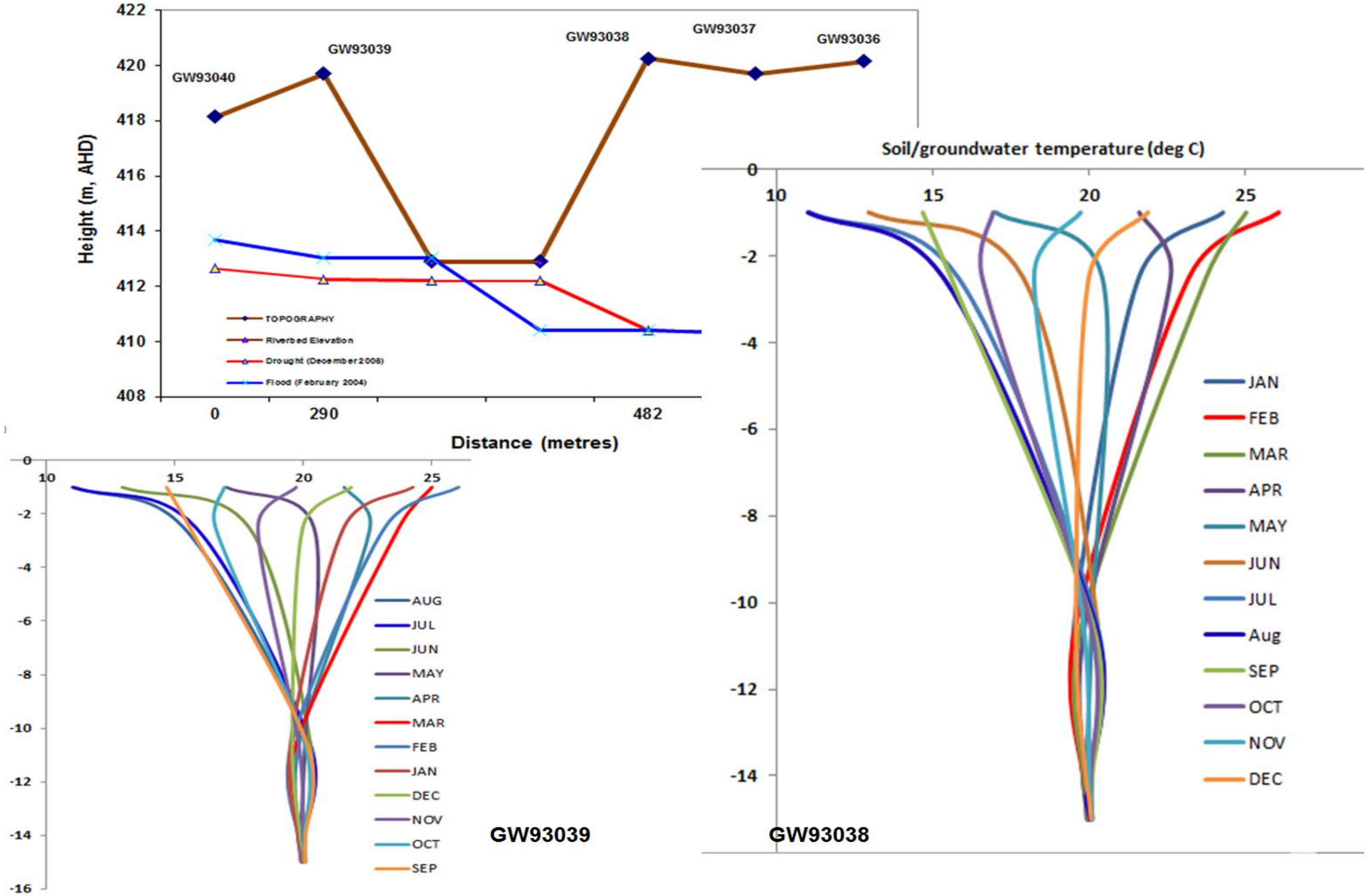


Figure 6-12. Measured temperature envelopes for the Kootingal transect. The monitoring bores (GW93038 and GW93039) are outside the thermal influence of the streambed region (a far field).

Table 6-3. Simulated monthly temperature profiles: Kootingal transect (under streambed, m)

Depth/Month	**1.92**	**2.46**	**2.83**	**5.2**	**9.22**	**14.1**	**19.3**	**24.2**
Dec-10	21.41	21.39	21.37	21.27	21.11	20.91	20.61	20.23
Jan-11	22.36	22.31	22.27	22.01	21.56	21.09	20.66	20.25
Feb-11	22.06	22.03	22.02	21.91	21.67	21.26	20.79	20.31
Mar-11	21.51	21.51	21.51	21.51	21.44	21.19	20.79	20.33
Apr-11	20.43	20.46	20.49	20.65	20.83	20.85	20.64	20.28
May-11	18.99	19.06	19.11	19.43	19.92	20.26	20.31	20.13
Jun-11	17.50	17.58	17.64	18.05	18.77	19.46	19.83	19.89
Jul-11	17.09	17.16	17.21	17.57	18.20	18.88	19.38	19.66
Aug-11	17.12	17.16	17.19	17.43	17.91	18.54	19.09	19.48
Sep-11	17.66	17.69	17.71	17.82	18.04	18.46	18.94	19.38
Oct-11	18.58	18.56	18.55	18.47	18.41	18.57	18.92	19.34
Nov-11	19.38	19.37	19.37	19.35	19.24	19.11	19.18	19.43

During the development of temperature envelopes under the streambed, measured streambed temperatures were available at 40 and 80 cm, below the streambed. As such the lower part of the profile is unconstrained by measured streambed temperatures at a lower depth below 80 cm. The envelopes are bounded by two extreme temperatures, in January and June. They seem to be similar with measured temperatures in GW93039 and GW93040. As temperature profiles in GW93038 and GW930039 are outside the influence of the streambed thermal zone, the temperature data collected in these two bores cannot be used for calibration. Nevertheless, the envelope patterns appear to be similar. The major difference is in the lower part of the profiles, where the measured temperatures may have been compromised by human interferences. The simulated fluxes at different depths are discussed in 6.6.10, where the effect of temporal aspects of scale is presented.

The pattern of thermal wave propagation depends on thermal diffusivity of the geological or soil layers. During the simulations, the thermal properties were kept constant; the only calibration parameter used was the hydraulic conductivity of the surficial sediments. The simulated temperature envelopes are similar to those reported in literature (Anderson, 2005; Bartolino and Niswonger, 1999a; Lapham, 1989a). Additional information on simulation procedures and ranges of estimated hydraulic conductivities is presented in Chapter 3 and Chapter 4 of this thesis.

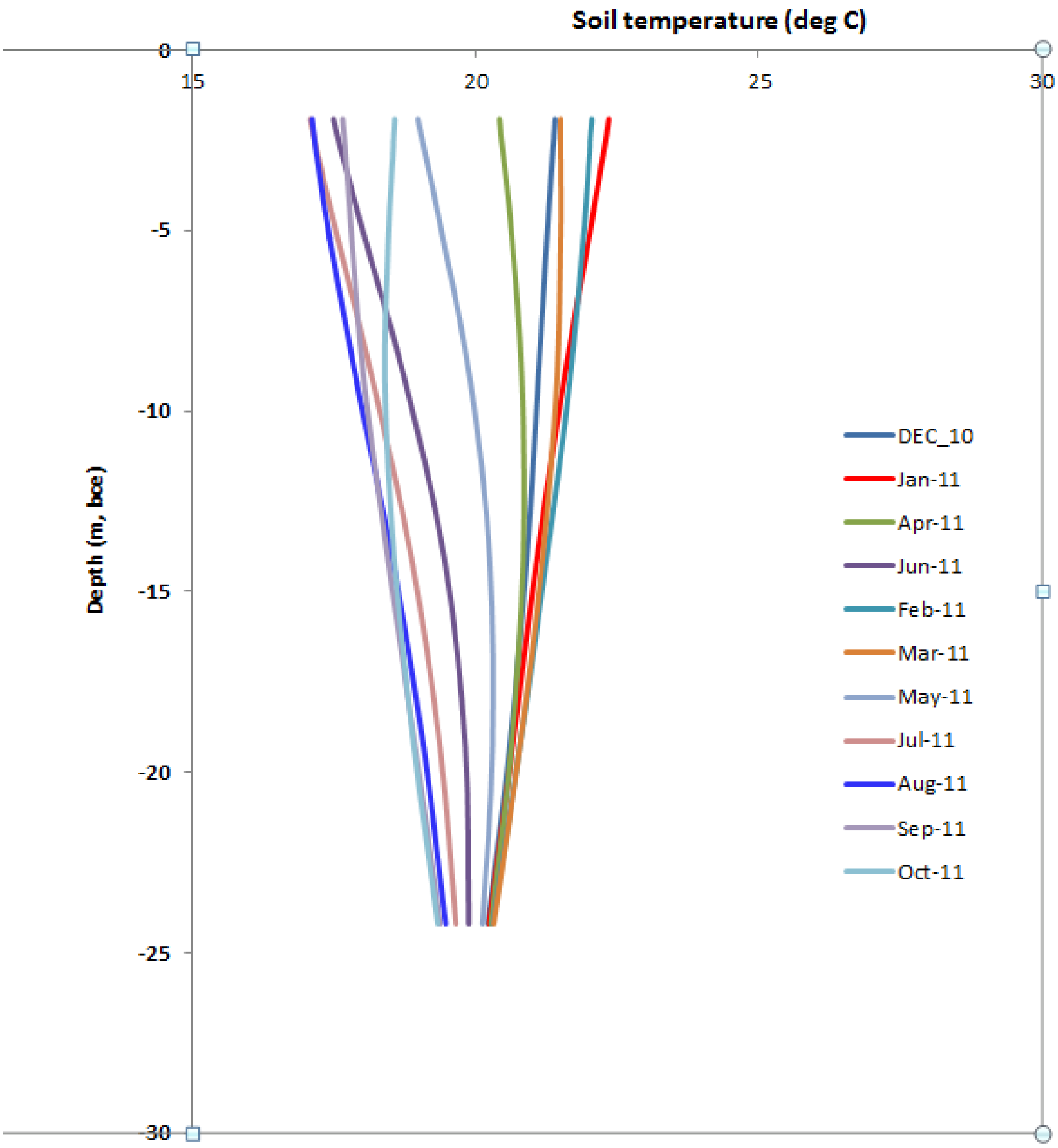


Figure 6-13. Temperature envelopes for 2011: Kootingal transect.

Simulated groundwater temperatures profiles are shown in Figure 6-13. There is a phase lag of about 6 months between surface and groundwater temperature signals. The phase lag is dependent on the thermal and hydraulic properties of the geological medium. In addition, from the simulated monthly temperature snapshots shown in Figure 6-14to Figure 6-26, it is obvious that the lateral propagation of thermal wave doesn't extend for more than 20 meters.

A more detailed seasonal analysis of thermal propagation in the streambed/phreatic zone for 2010/11 is provided below.

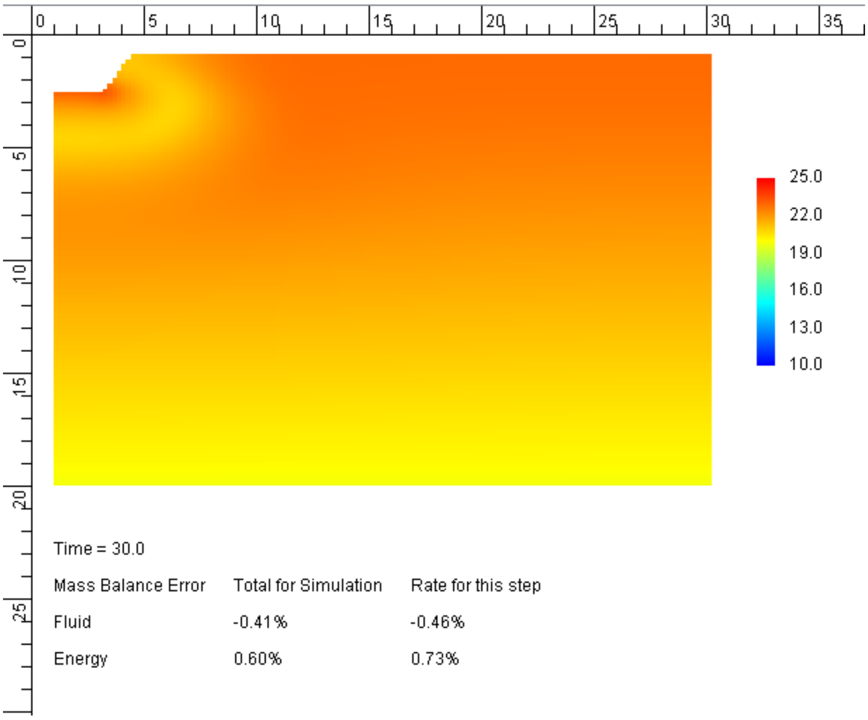


Figure 6-14. Propagation of heat waves within the streambed and phreatic zone at a monthly scale: Dec 2010

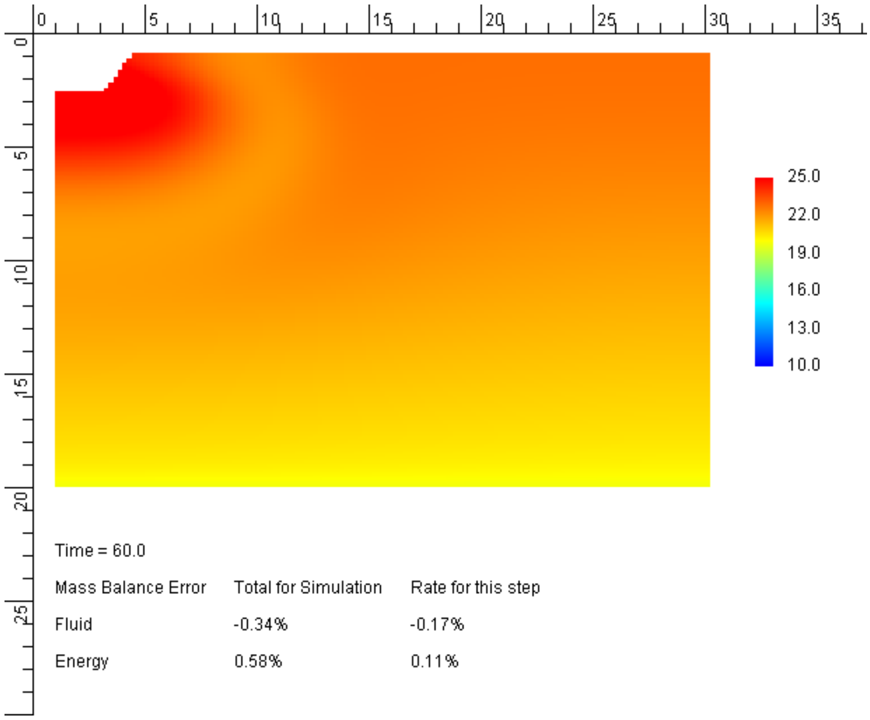


Figure 6-15: Propagation of heat waves within the streambed and phreatic zone at a monthly scale: January 2011

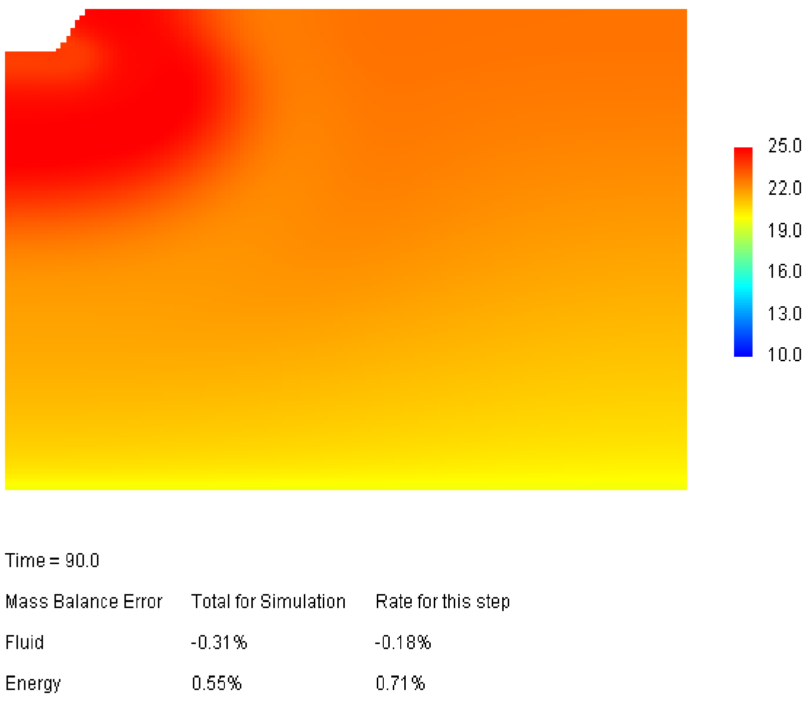


Figure 6-16 Propagation of heat waves within the streambed and phreatic zone at a monthly scale: February 2011

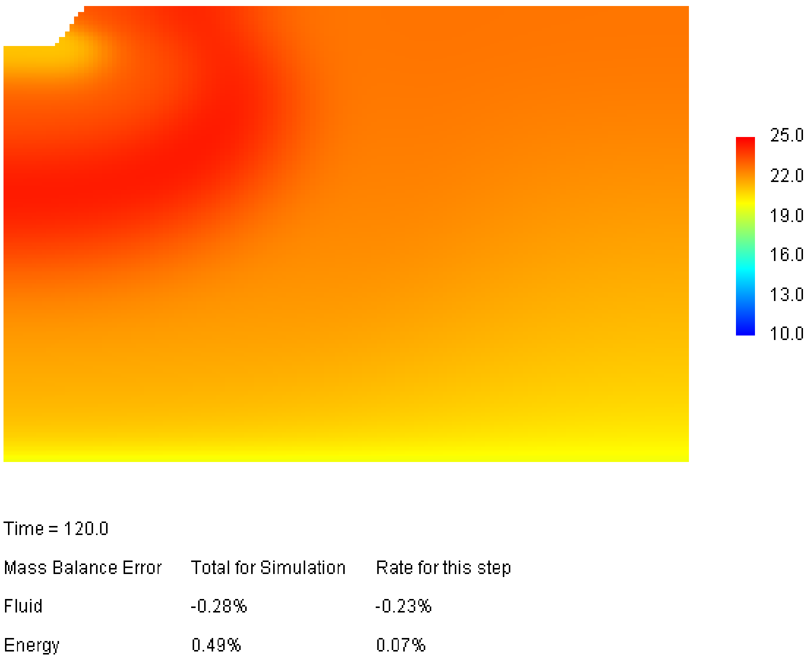


Figure 6-17. Propagation of heat waves within the streambed and phreatic zone at a monthly scale: March 2011

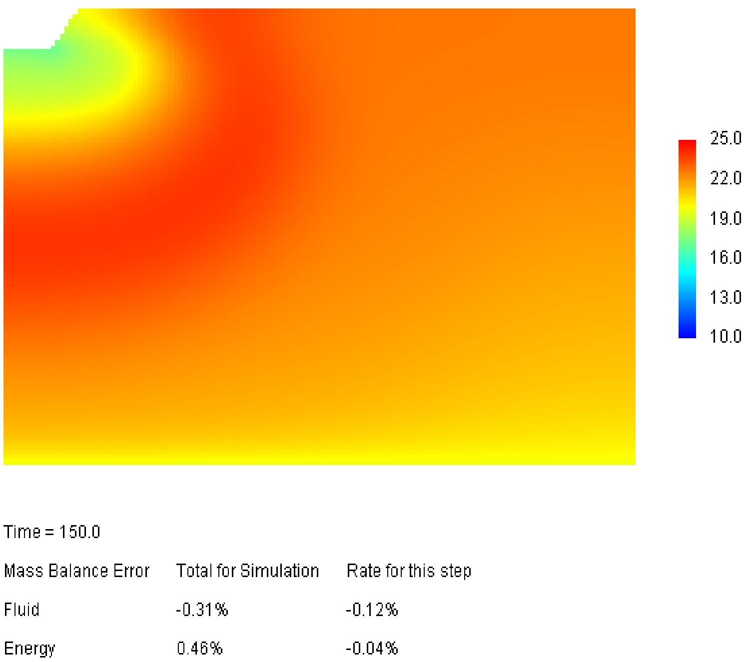


Figure 6-18. Propagation of heat waves within the streambed and phreatic zone at a monthly scale: April 2011

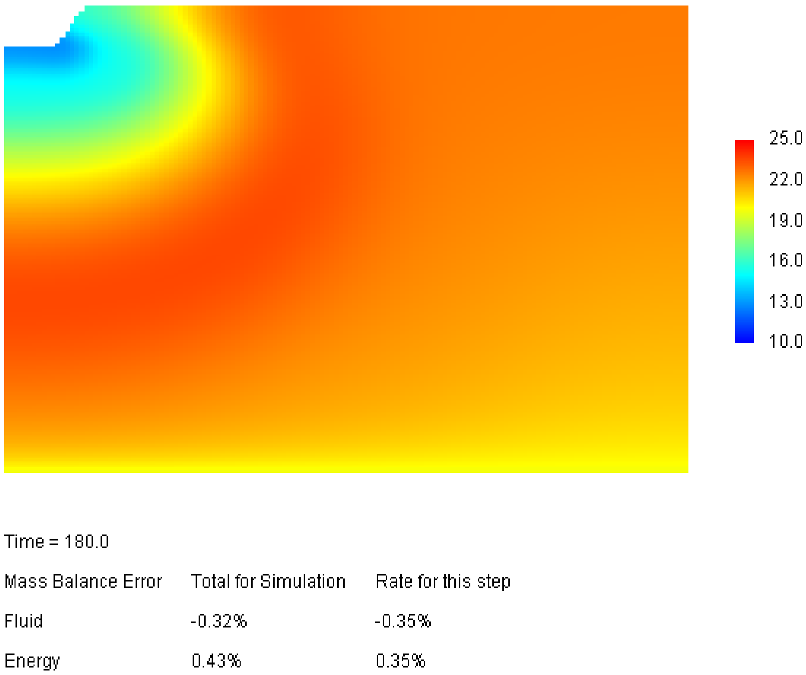


Figure 6-19. Propagation of heat waves within the streambed and phreatic zone at a monthly scale: May 2011

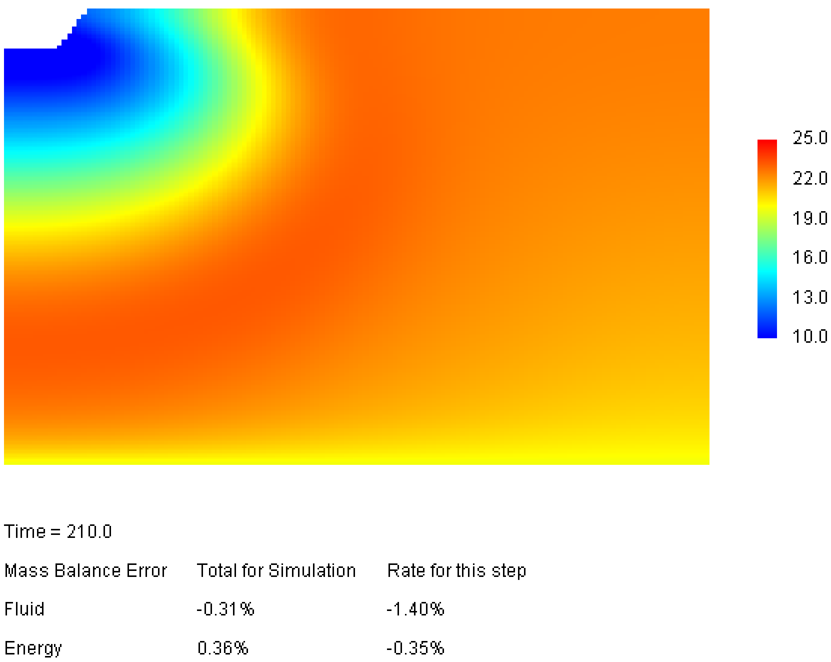
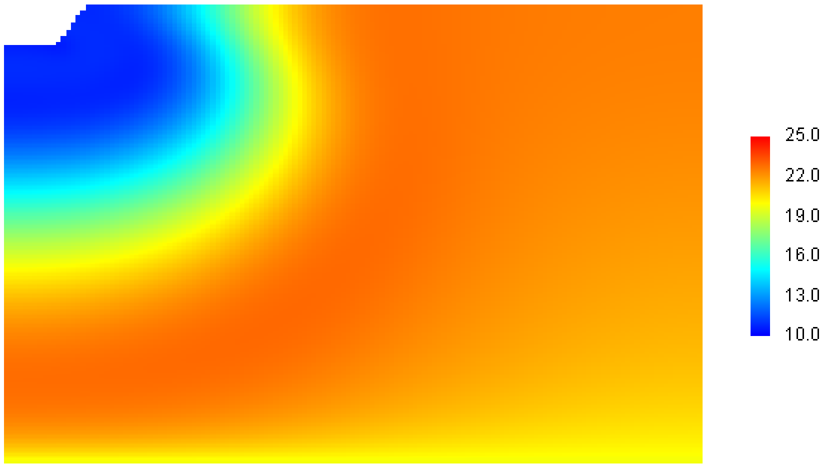


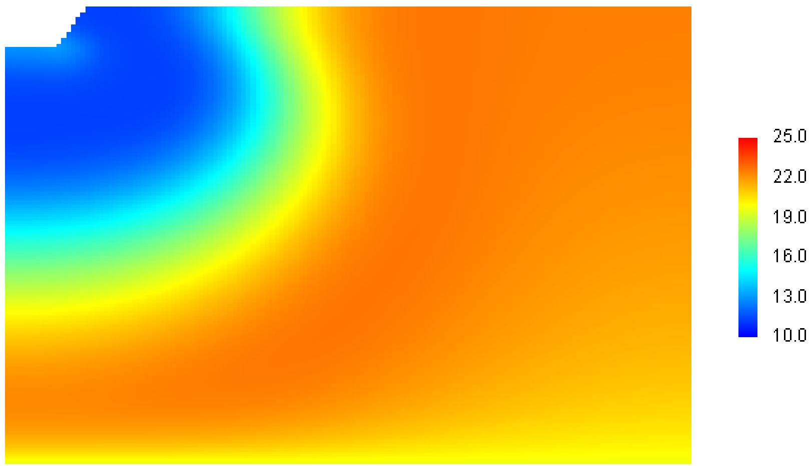
Figure 6-20. Propagation of heat waves within the streambed and phreatic zone at a monthly scale: June 2011



Time = 240.0

Mass Balance Error	Total for Simulation	Rate for this step
Fluid	-0.32%	-0.21%
Energy	0.29%	-0.30%

Figure 6-21. Propagation of heat waves within the streambed and phreatic zone at a monthly scale: July 2011



Time = 270.0

Mass Balance Error	Total for Simulation	Rate for this step
Fluid	-0.28%	0.59%
Energy	0.21%	-0.71%

Figure 6-22. Propagation of heat waves within the streambed and phreatic zone at a monthly scale: August 2011

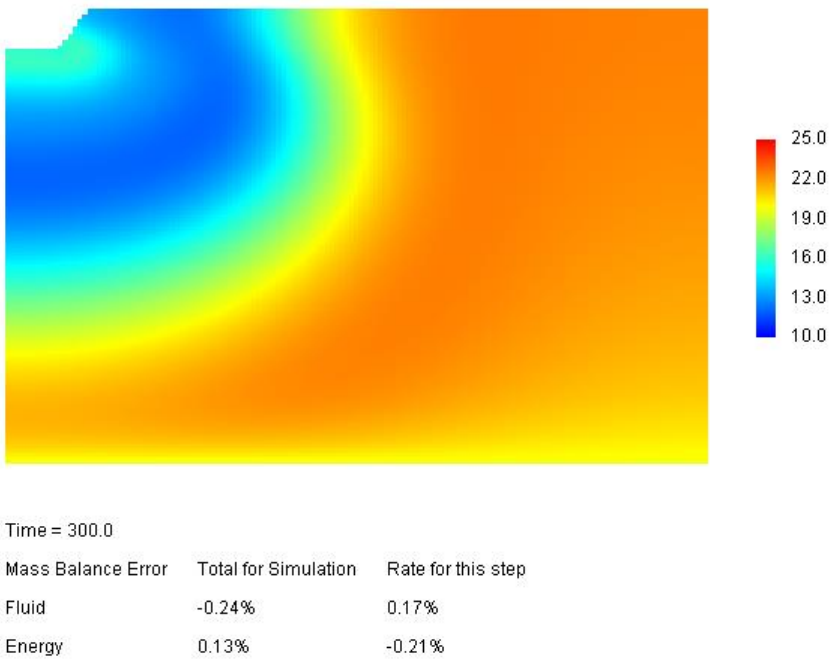


Figure 6-23. Propagation of heat waves within the streambed and phreatic zone at a monthly scale: September 2011

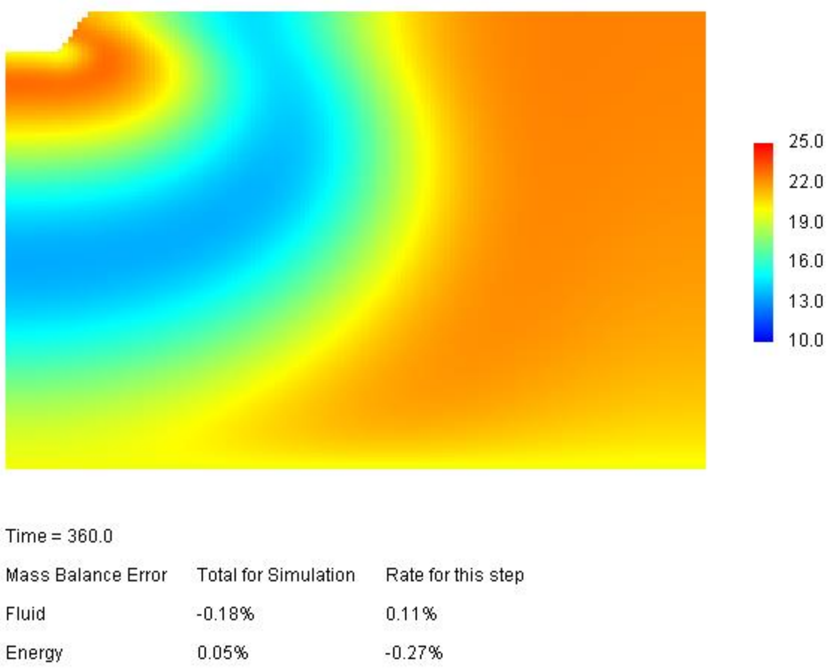


Figure 6-24. Propagation of heat waves within the streambed and phreatic zone at a monthly scale: October 2001

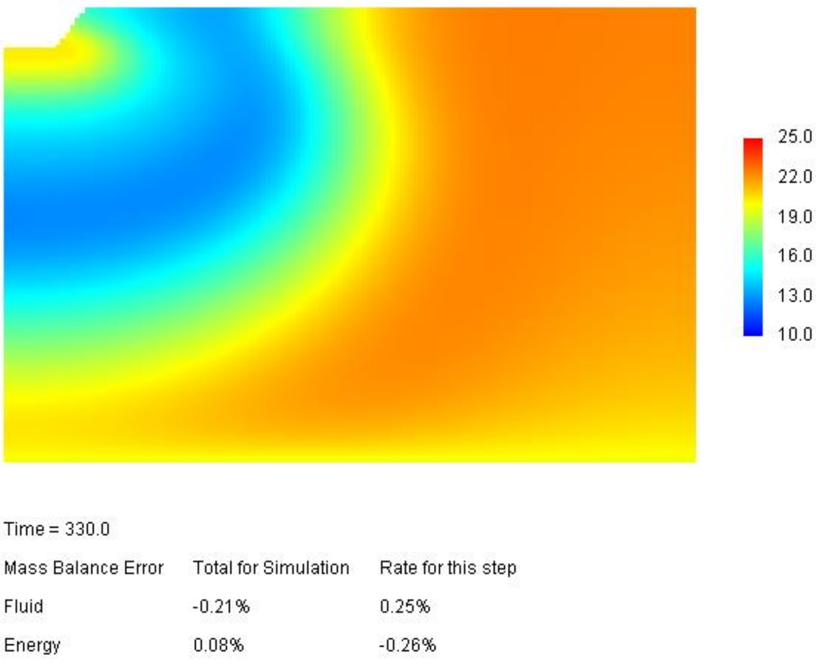


Figure 6-25. Propagation of heat waves within the streambed and phreatic zone at a monthly scale: November 2001

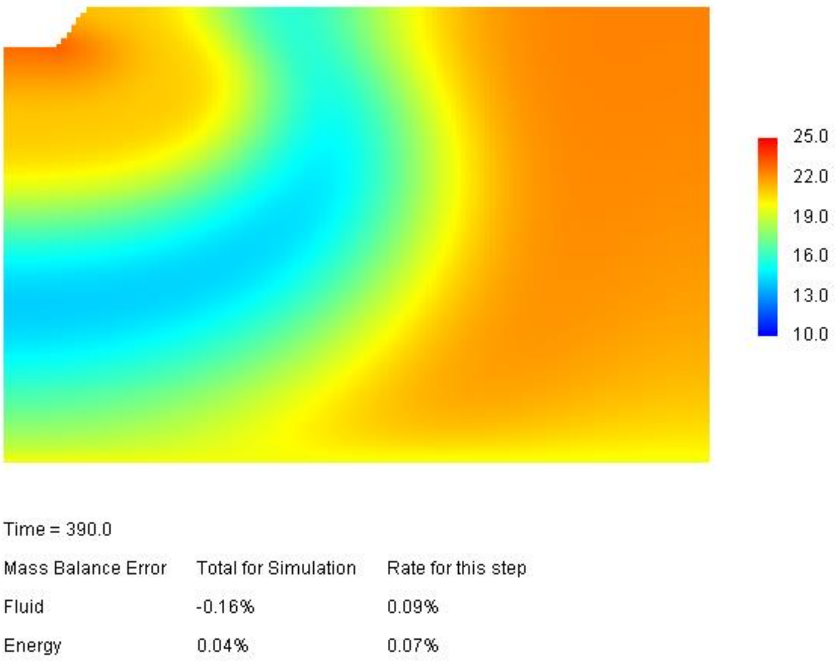


Figure 6-26. Propagation of heat waves within the streambed and phreatic zone at a monthly scale: December 2011

Based on the above seasonal analyses of bore temperature profiles, the Kootingal transect is a predominantly a losing system.

6.6.6.2.1 Summer streambed thermal regime (December 2010-February 2011)

A full 13 months of thermal data from 1 December 2000 to 31 December provided an opportunity for more detailed analyses of surface-groundwater connectivity issue. For instance, during winter 2010/11, stream stages and temperatures at the Kootingal transect varied from 0.36 to 2.50 m and 17.86 to 28.6^{o}C respectively. In sympathy with the upper boundary conditions, streambed temperatures varied from 17.65 to 2.85 ^{o}C.

Abrupt rises in riparian zone water table (as illustrated in Figure 4-13 and Figure 4-6) are associated with high runoff and flood events. Based on Darcy's equation, these events induce high hydraulic gradients between surface water in the channel and groundwater. Thus, combination of high runoff and high temperature create a favourable environment for high seepage rates. For summer season of 2010–11, the calculated infiltration rate ranged from 0.465 to 0.693 m/d.

6.6.6.2.2 Autumn thermal regime (March 2011 to May 2011)

A transition of weather started to take place in autumn; during this period stream stages and temperatures at the Kootingal transect varied from 0.23 to 1.2 m and 11.5 to 24.7 C respectively. The highest stream stage observed in autumn is by 50% lower when compared with the previous season. Relatively colder stream water started to propagate into the system, inducing a gradual decrease in streambed temperatures. The propagation of a cold front, induced both by conduction and advection, continued until the end of September 2011. During autumn, the simulated infiltration rate ranged from 0.437 to 0.595 m/d.

6.6.6.2.3 Winter thermal regime (June 2011 to August 2011)

In winter the minimum stream minimum observed stream temperature dropped to 7.6 ^{o}C and stream stage varied between 0.48 to 1.7 m. Persistent low stream temperatures in combination with relatively low stream stage did create a favourable environment for stream induced infiltration. Historical stream flow data suggest that flooding is more frequent in summer and spring, but take place in winter. During winter, the simulated infiltration rate ranged from 0.403 to 0.487 m/d.

6.6.6.2.4 Spring thermal regime (September 2011 to November 2011)

In spring, stream stages and temperatures varied from 0.46 to 3.85 m and 7.6 to 25.6ºC respectively. After the onset of spring 2011, stream water temperatures at the Kootingal transect started to rise, causing a reversal in of streambed temperatures within the streambed zone (Figure 6-10); the simulated flux ranged from 0.432 to 0.712 m/d.

6.6.7 Sub-catchment scale

At a sub-catchment/catchment scales, recharge discharge mechanisms may be better conceptualised using the principle of groundwater flow systems. Sub-catchment scale interaction between surface and groundwater is represented by a longitudinal hydrological cross section, extending between Mulla Mulla and Kootingal gauging stations (Figure 6-27). Although the density of monitoring bores in the upland catchment is sparse, the relative position of the water table in relation to the surface topography appears to be lower. At a sub-catchment scale, the surface water recharges the groundwater system. However, in some cases depending on the streambed topography, localised upwelling and downwelling may occur.

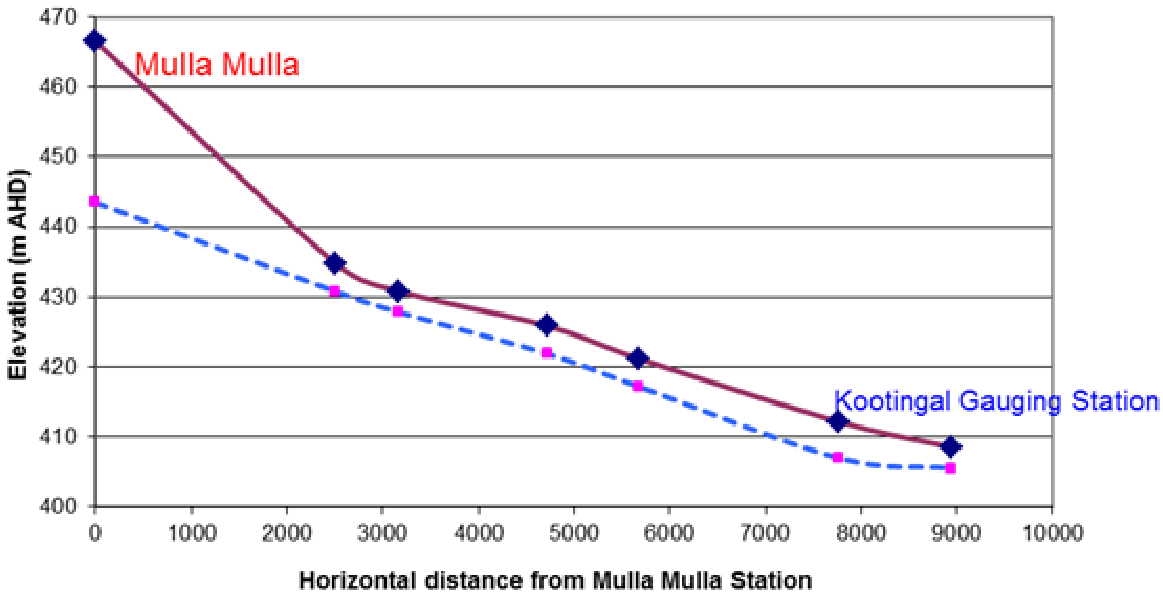


Figure 6-27. Hydrological length-section between Mulla Mulla and Kootingal

As reported earlier (Berhane et al., 2008), the Cockburn River losses substantial volumes of water between Mulla Mulla and Kootingal gauging stations. The losses are both natural and human induced. For example, close to the Mulla Mulla gauging station, surface water extraction takes place. No attempt was made to differentiate between natural losses and pumping induces losses.

Although not part of this PhD study, low flow analysis along the different reaches of the Cockburn Valley was carried out by Parson et al. 2006 (NSW Department of Natural Resources; see Figure 6-28 to Figure 6-33). This hydrologic survey demonstrates that the Cockburn Valley is a losing stream at catchment scale. In fact, the Cockburn River ceases to flow at the Nemingha Bridge on 10th August 2006 (Figure 6-33).

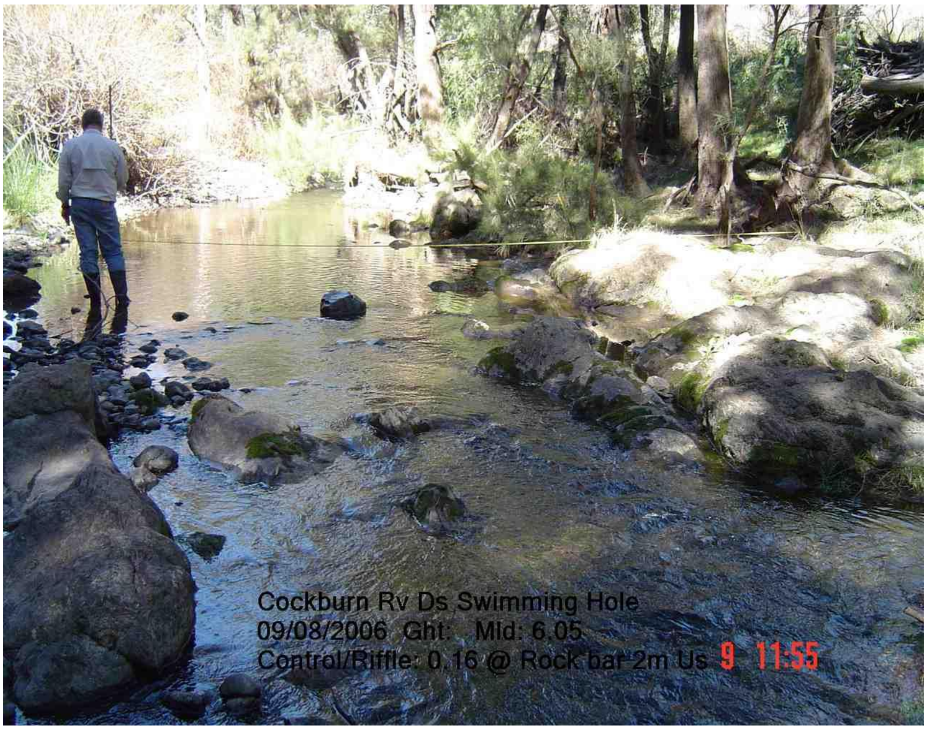


Figure 6-28. Stream velocity measurements downstream of swimming hole

Figure 6-29. Stream velocity measurements using current meter near the Chapmans well

Figure 6-30. Stream flow estimation at Sandy Creek, Kootingal

Figure 6-31. Stream velocity measurements using current meter at the Kootingal Bridge

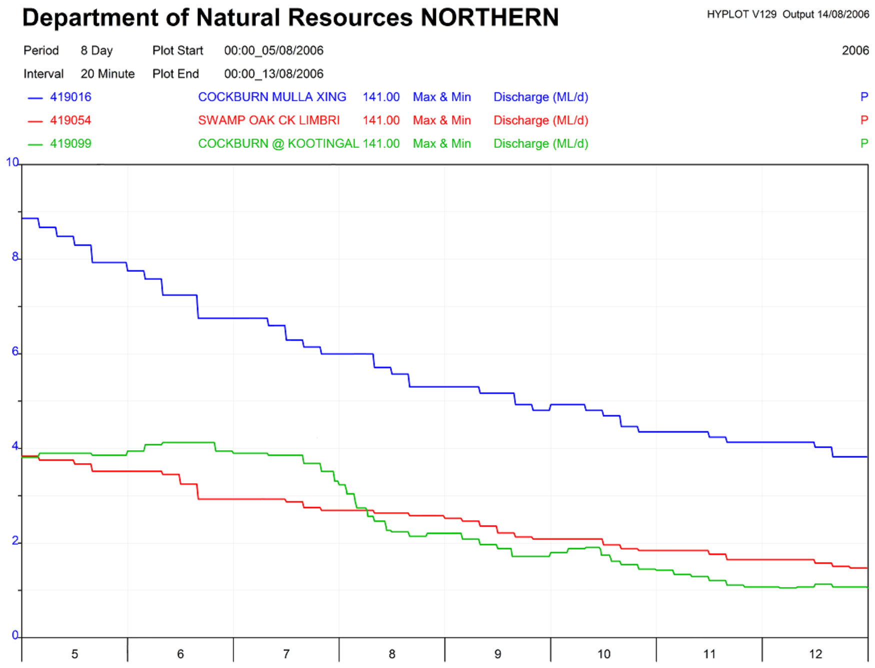


Figure 6-32. Reported stream discharges at three reaches along the Cockburn Valley in 2006.

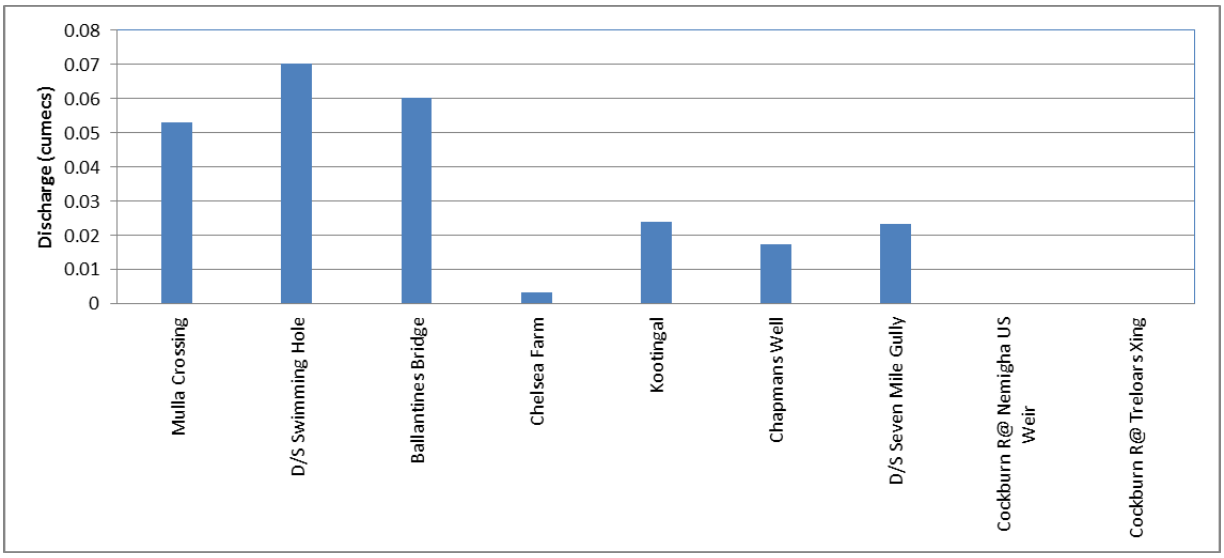


Figure 6-33. Variation in stream discharge between Mulla Crossing and the Nemingha Bridge (Parsons et. al, 2006)

6.6.8 Water exchanges at catchment/regional scale

To my knowledge, with the exception of the Great Artesian Basin, the regional and 'intermediate' groundwater systems in the Namoi Valley do not contribute to stream discharge (Chapter 2 and Chapter 3 of this thesis). In contrary, the stream drainage systems in the Namoi Valley act as a sink to the surface water systems.

On the basis of hydrogeological, geomorphological and structural factors, three main types of hydrogeothermal zones can be recognised: closed, open and combined. In this study, an open system is the focus, where surface water interacts with the phreatic aquifer system. Outside the alluvial groundwater management areas, the groundwater systems in the Namoi Valley can be considered as closed systems, due to low magnitude of recharge.

The influence of changes in temperature from the surface propagates into the subsurface at a rate determined by the diffusivity of the geological or soil layers and the depth of penetration depends on the amplitude and duration of the perturbation. We should be aware that both the paleoclimate and seasonal effects are important to consider when collecting down-hole temperature measurements as the groundwater bores used have total depths which fall within the zone of influence (< 500 m) (Danis, 2014).

Although bore temperature profiles, representing contrasting geological environments in the Namoi Valley are not readily available, for illustrative purposes, an attempt has been made to compare bore temperature profiles from two different geological environments. In this example, the first case study represents a bore temperature profile from a shallow groundwater system in the Maules Creek catchment (open system), while the second case study represents a closed groundwater systems in the Gunnedah Basin (closed system), which is considered to have been recharged in the geological past. Based on visual inspection, the bore temperature profile in GW967137 (Maules Creek Catchment) is similar to bore NC112 (Gunnedah Basin) and is shifted by about 2°C to the warmer side of the temperature scale. At this stage, it is not clear whether the shift in groundwater temperature profiles in the two contrasting geological environments is caused by the recent drought or climate change.

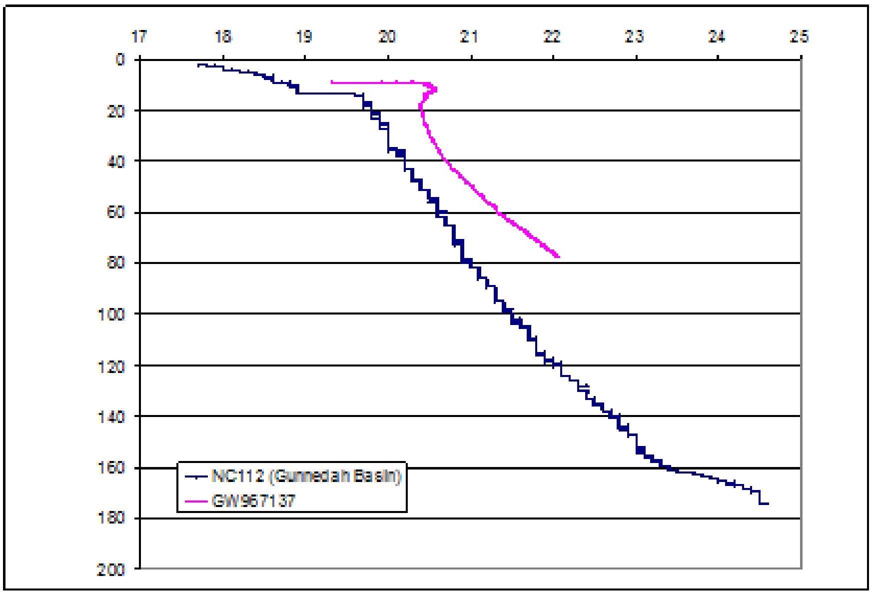


Figure 6-34. Bore temperature profiles representing an open system (Namoi Alluvium) and a 'closed' geological system overlain predominantly by aquitards (Gunnedah Basin/ GAB Sediments).

Recent hydrological studies carried out in the Namoi Valley add value to our understanding of timing and duration of recharge processes in ephemeral river systems. The phreatic groundwater system in the Maules Creek Catchment is recharged episodically by stream losses of the Maules Creek (Figure 6-14 to Figure 6-26). Consequently, the temperature profile in GW967137 is responsive to temperature oscillations on the surface. In contrast, the Gunnedah Basin is dominated by palaeowaters from earlier recharge episodes (Chapter 2), and possibly embeds temperature signals from the past. Thus, the temperature profile measured in NC112 is probably not influenced by modern recharge and reflects conductive geothermal conditions (Figure 6-34). Given the paucity of hydrological information over a wide area in the Namoi Valley, the interpretation should not be taken at face value.

At long time scales, deep groundwater systems do not change in a dramatic fashion. In contrast, the phreatic groundwater system is sensitive to episodic recharge events, where a sudden flux of water from a river into the shallow groundwater system induces a new heat balance, hence causing the groundwater temperature to increase or decrease depending on the signature of river water temperature. If the river water temperature is greater than the groundwater temperature, a positive anomaly is produced. Vice versa, a negative groundwater temperature anomaly is observed. Thus, freshly infiltrated (hyporheic/streambed) water can be distinguished from the ambient groundwater by the dilution effect in EC and temperature and its short residence time in the subsurface of up to a few days.

In the context of this study, thermal anomalies are associated with stream induced recharge in ephemeral and intermittent streams, when the slowly changing seasonal groundwater temperature is disrupted by a sudden flux of surface water. Here below, we provide three case studies, where the occurrence of thermal anomaly was observed in the Namoi Valley, thus demonstrating the high level of connections between surface water and groundwater in this area.

6.6.8.1 Negative thermal anomaly: Well-4

A negative anomaly was observed in an infiltration gallery (Well-4), located in the middle part of the catchment. The production well is the most productive among the five bores which supply the township of Kootingal. But it is also the first production bore to dry up, during extended dry spells. This particular bore installation captures water directly from the stream by a buried horizontal underground screen, connecting Well-4, directly to the Cockburn

River. There is therefore no doubt that in this case water fluxes are from the river to the aquifer (Figure 6-35).

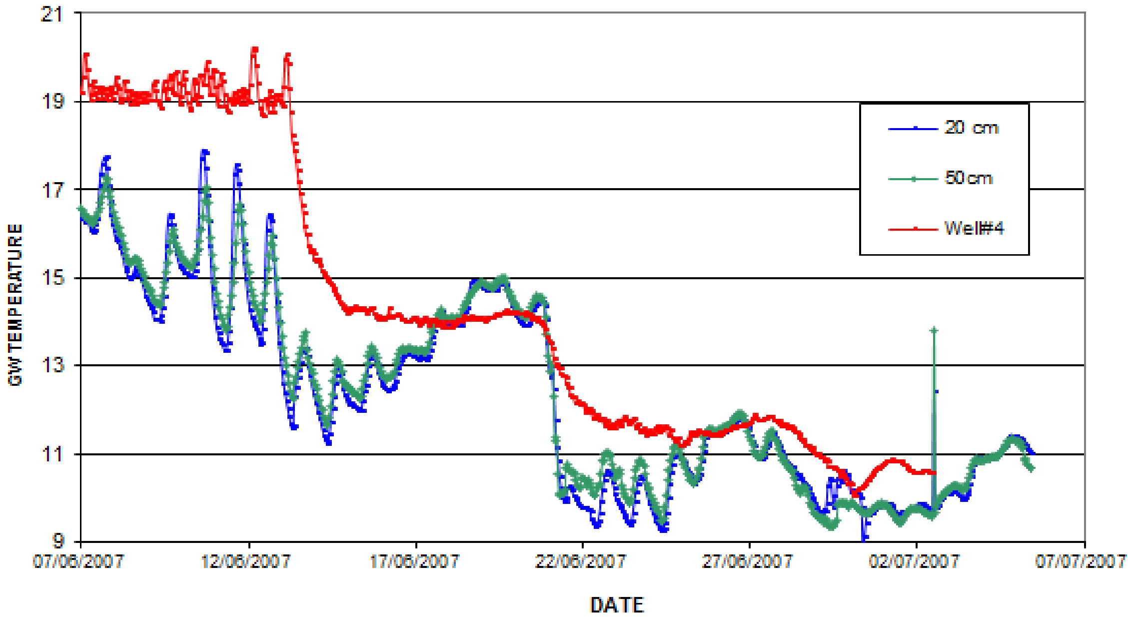


Figure 6-35. Sediment and groundwater thermographs in the vicinity of Well 4: Cockburn Valley. There were two thermistors at 20 and 50 cm depth and one installed in the well.

In some cases, runoff generated in the upper part of catchment may have an effect on the riparian groundwater systems located in the lower part of the catchment. This particular event was observed in June 2006, where the groundwater temperature in Well 4 dropped from 19 to 14 °C within a three day period (Figure 6-35). The total cooling effect or “negative” anomaly that took place during injection of water into the channel was estimated using the following simple heat balance equation (Lancaster and Haggerty, 2005):

$$Energy = C \Delta T \rho_w V_C \qquad \text{Eq. 6-12}$$

where C is the specific heat of water (4186 J/°K-kg), ΔT is side channel’s change in temperature (19 to 14 °C), ρ_w is the density of water (998.29 kg/m3), and V_C is the volume of the channel (100 m^3). Applying the above values to equation (Eq. 6-12), the total cooling effect due to the infiltration gallery (channel effect) is 1.25×10^9 J.

6.6.8.2 Positive and negative thermal anomalies: Maules Creek

In a similar fashion a positive anomaly was observed at in GW273033, piezometer located close to the Nemingah Bridge, near Tamworth. This particular site is downstream of the confluence of the Peel River (regulated) and the Cockburn River (unregulated). On a high runoff event that occurred on 1st September 2008, the groundwater level rose by about 0.45m. The sudden influx of relatively warmer surface has induced a change in groundwater temperature by about 0.08°C (Figure 6-36). Assuming a specific yield of 0.2, this particular hydrologic event generated a point recharge of 0.09m. Up-scaling these point recharge estimates to a catchment scale is a challenging exercise, however it once again clearly demonstrates the existence of vertical downward fluxes rather than flow from the groundwater compartment into the river.

A more dramatic change in groundwater temperature was observed in one of the NSW Office of Water piezometers in Maules Creek (GW967138), where a change in groundwater temperature up to 2°C was observed during high runoff events in the summer of 2006. In contrast, a runoff event that occurred on 2nd January 2010, generated a negative anomaly with a change in temperature by about 0.1°C. Depending on stream temperature signal, a negative or positive temperature anomaly can occur at the same site on different times. However, both indicated vertical downward fluxes (Figure 6-37).

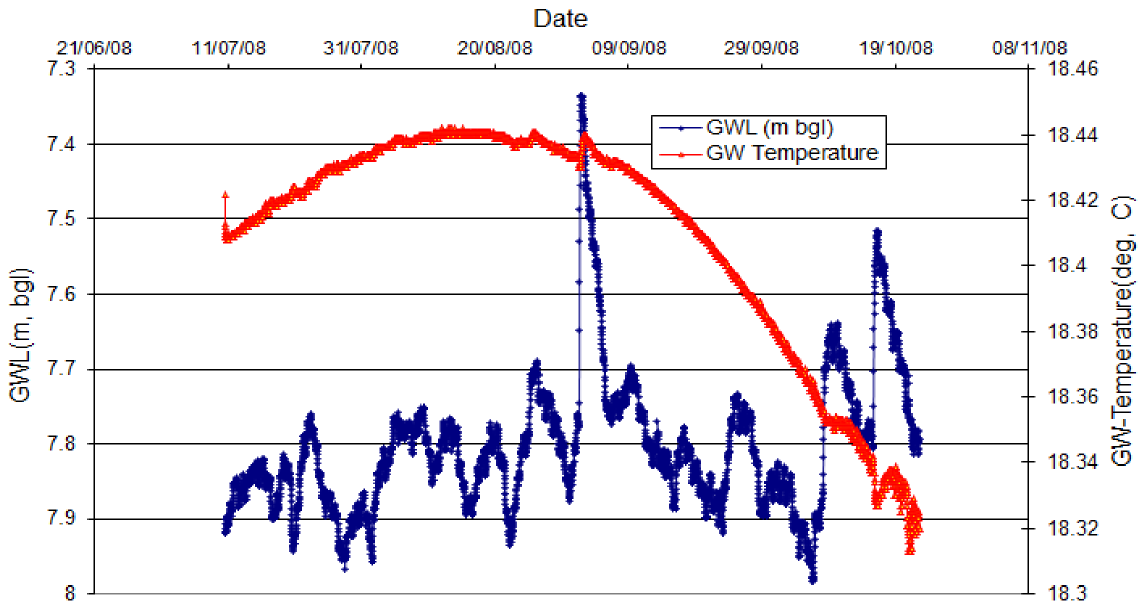
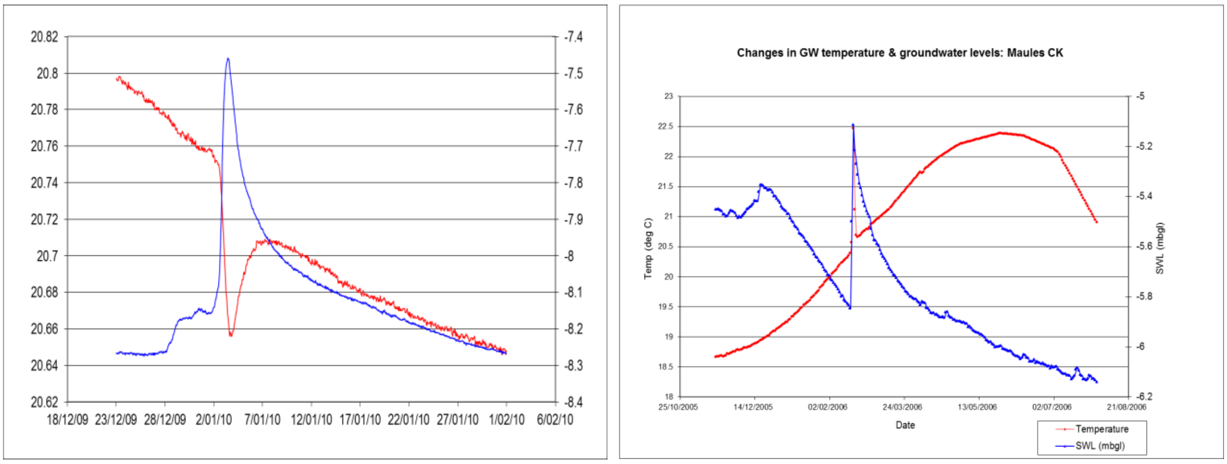


Figure 6-36. A positive thermal anomaly observed in GW273033: Nemingha Bridge



A: Positive anomaly

B: Negative anomaly

Figure 6-37. A positive and negative thermal anomaly observed in GW967138: Maules Creek Catchment. The hydrothermic information is providing vital information on timing and frequency of groundwater recharge in ephemeral and intermittent river systems.

6.6.9 Simulated infiltration rates at daily and monthly stress periods

From a catchment hydrology sampling point of view, found that daily and or monthly data miss much information and argued that high-frequency measurements of chemical behavior will provide new insights into subsurface storage of water within catchments and the flowpaths by which water reaches the stream. In a similar fashion, timing and frequency of ephemeral groundwater recharge can be assessed by designing a compatible high frequency surface and groundwater monitoring network.

Monthly time steps are usually used to mimic hydrological processes at catchment and sub-catchment scales in humid environments. However, for estimation of recharge/infiltration rates of ephemeral river systems, a monthly time step may not necessarily capture episodic runoff events. If it were not for high frequency sampling of groundwater levels and temperatures, the negative and positive anomalies discussed above would not have been captured (Figure 6-36 and Figure 6-37) for the study reaches. Thus, the effect of temporal domain discretization on infiltration rates was evaluated using an hourly and daily stress periods for the Kootingal transect.

Multiple hydrologic events, with stream stage ranging from 0.26 to 2.7 m were captured in 2011 and hydrothermic data set from this year was used to: (1) assess the effect of time discretization on the simulated fluxes; and (2) study the variations of flux with depth (5.4.1).

The simulated fluxes for hourly and daily stress periods are shown in Figure 6-38. It can be seen that the selection of stress periods on simulated fluxes is significant. Simulated flux using an hourly stress period shows up to 40% increase, in fluxes when compared to a daily stress period. It should be noted that daily data was obtained by aggregating hourly data. During this application, the high frequency temperature and water level fluctuations are smoothed. This procedure might have biased the outcome of this comparison using different stress periods.

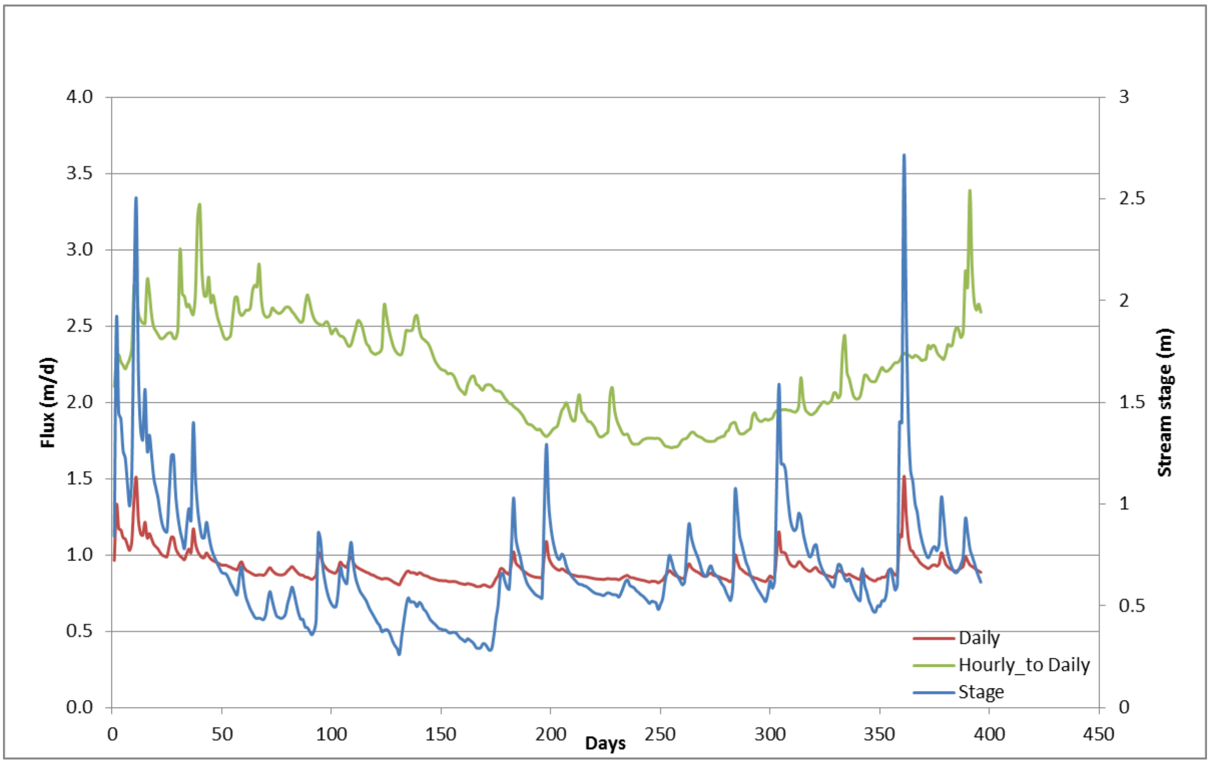


Figure 6-38. Simulated fluxes at different for an hourly and daily stress periods: Kootingal transect

6.6.10 Simulated fluxes at different depths (from infiltration to recharge)

In terms of the temporal frequency of the occurrence of surface runoff, 2011 was the wettest year, since the installation of streambed thermistors at the Kootingal site in April 2008. Thus, this particular was selected to study the variation of infiltration with depth. During the simulation, the stream stage and temperature representing the upper boundary condition was allocated measured daily values, while the lower boundary condition was allocated a fixed groundwater temperature of 19.8 °C and a head of -5 m bcl. For comparison purposes, infiltration rates were simulated at depths (2, 2.49, 5.42, 9.18, and 14.2 m), with a reference channel elevation of 1.8 m. At depth of 2 m (0.2m bce), infiltration rate varied from 0.4 m to 0.76 m /d, while at depth of 14 m (12.2 m bce), infiltration/recharge varied from 0.09 to 0.15

m/d, showing a smaller range of variation on a yearly scale. As shown in Figure 6-39, infiltration decreases with depth. The attenuation of infiltration rate with depth can be caused by wetting of an unsaturated soil, lateral flow and evapotranspiration.

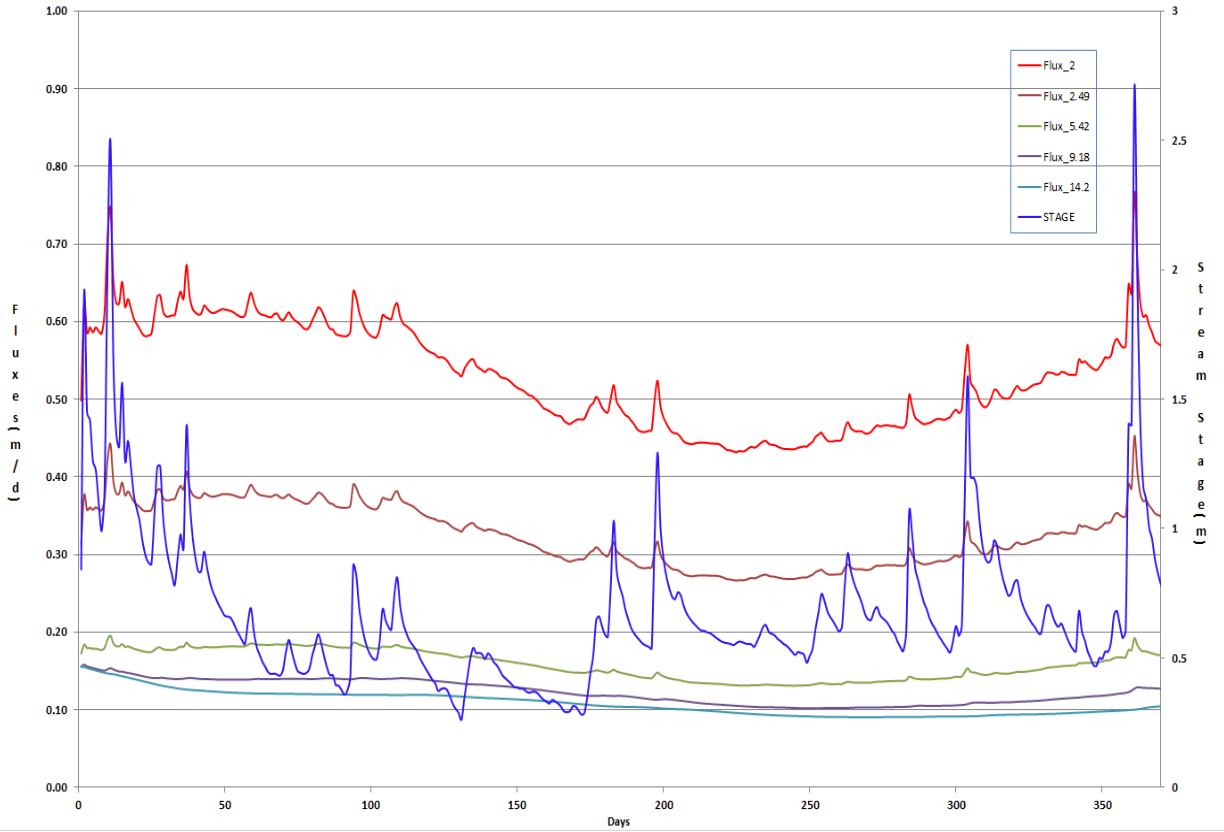


Figure 6-39. Simulated infiltration, percolation and recharge rates for 2011: Kootingal transect.

The estimated infiltration rate is in the same order of magnitude compared to infiltration rates reported for study sites in the US (Constantz et al., 2002) and Namibia (Dahan et al., 2008).

6.6.11 DiscussionLimitations of the Bredehoeft Method

The Bredehoeft method (1965) was originally applied for a basin scale study, when due the size of the of the study area a steady state condition can be assumed. More recently, this method has been applied for a small scale investigation in the hyporheic zone. However, temperature profiles beneath streambeds can be seasonally transient or steady-state (Anibas et al., 2009), and a steady-state approach by Bredehoeft and Papadopulos (1965) may result in inaccurate estimation of groundwater flux for seasonally transient profiles. Therefore, the calculated fluxes shown in Table 6-1 are considered to be more reliable for the RHS of the bank, where a relatively stable groundwater discharge takes place. In contrast, the losing section of the Kootingal transect the variability of flux is influenced by multiple factors (diurnal stream temperature variations, stream stage, presence or absence of the clogging layer). In addition, the meaning of *L* included in the Peclet ratio (Equation 3) is unclear from the literature (Silliman et al., 1995). Does *L* represent a sampling depth or a thickness of a geological layer?

6.6.11.2 Near streambed interactions across the multiple scales

Toth (1963), demonstrated how steady-state groundwater flow fields develop in response to an assumed relationship between topography and the groundwater surface. His conceptualisation of regional groundwater flow has proved particularly robust and has been regularly used in the last 5 decades. Recently, Toth's methodology been has applied to visualize water flow patterns at a streambed scale (Stonedahl et al., 2010). 'Hyporheic' flow has been conceptualized to occur at three spatial and temporal scales (Figure 6-40). At the finest scale (streambed scale), hyporheic flow is driven by alternating pool/riffle sequences in stream channel (Vaux, 1968). Streambed scale pathways may be anywhere from 10^{-2} to 10 days in duration.

At an intermediate spatial scale (meander-bend scale) 'hyporheic' flow is driven by the development of mid-channel bars and meander bends (Wroblicky et al., 2010) by the presence of side channels, backwaters, and abandoned channels (Stanford and Ward, 1993). Streambed scale pathways may be anywhere from 10 to 1000 days in duration.

At the coarsest scale (floodplain scale) water tends to enter the alluvial aquifers at the upstream end of flood plains, flow laterally through the alluvial aquifer, and re-emerge the lower end of the floodplain (Stanford and Ward 1993). Therefore, floodplain scale hyporheic flow arguably may be better conceptualised as more ‘classic’ aquifer recharge/discharge dynamics (Stonedahl et al., 2010). The complexity of the system under study which is clearly demonstrated from the results in this chapter means that we are still far from understanding the overall exchange between surface and groundwater. However, what emerges is that depending on the scale of observation different processes are either amplified or smoothed out. Future research should evaluate how this might influence management decisions, mostly taken at larger scales.

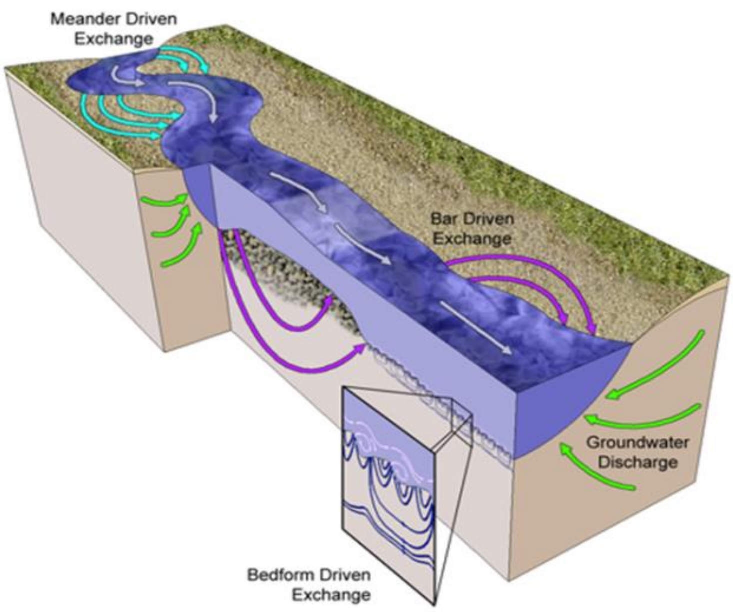


Figure 6-40. Near streambed interactions across the multiple scales (Stonedahl et al., 2010)

6.7 Summary and conclusions

The conclusions for this section are:

- Although drilling under-streambed can be an expensive endeavour, bore temperature envelopes are useful tools to delineate reaches into gaining and losing reaches;
- A temperature T-bar deployed in this project has been useful for mapping small scale streambed features. However, during the walkover survey conducted in June 2010, we might have destroyed small scale features;
- Due to erosion, thermal processes for small scale features (dunes and anti-dunes) are difficult to measure and map in a field environment;
- The recently revamped surface and groundwater monitoring network in the Namoi Valley provided vital hydrological information on timing and duration of recharge in intermittent and ephemeral river systems;
- Diurnal temperature oscillations are not observed below 0.5m depth in dry soils, whereas seasonal heat waves can penetrate up to 15m depth below the surface (Figure 6-13) in groundwater. The depth of penetration depends on the thermal diffusivity of the ground and water;
- Below the influence of surficial zone (15m bGL), groundwater temperatures are controlled by the local thermal gradients. In the Cockburn Valley, at a depth of ~15m the temperature is about 20ºC. However, in river reaches, which exhibit episodic recharge events, the slowly changing seasonal groundwater temperature is disrupted;
- During the propagation of a heat wave through a porous medium, induced by conduction and advection, reduction and shift of phase in depth occur. In the case of

the Peel Valley, the surface and groundwater temperatures are out of phase by almost six-months;

- Simulated monthly infiltration rates decrease with depth. The unaccounted water is consumed by evapotranspiration, lateral flow and wetting dry sediments in the unsaturated zone;
- Monthly stress periods are usually used to mimic hydrological process at catchment and sub-catchment scales in humid environments. However, for estimation of recharge/infiltration rates of ephemeral river systems, a monthly stress period may not necessarily capture episodic runoff events, that can be important for management;
- For reach and streambed scale investigations, a sub-hourly stress period is recommended. Daily and higher stress periods don't capture important hydrological events;
- At sub-catchment and catchment scales, a nested groundwater flow system as suggested by Toth was not observed in this study. Importantly, a bulk of recharge takes place along depression and creek lines. The occurrence of a positive head, which contributes groundwater discharge to streams is a rare phenomenon in arid and semi-arid environments. However, bank storage stream discharge, preferential path ways discharge may occur during extended runoff-events.

6.8 Recommendations

The recommendations for this section are:

- Because of the density of gauging station and availability of historic hydrologic data sets, the Cockburn Valley is an ideal The Cockburn Valley to test hypothesis and study intermittent stream characteristics in NSW. Thus, the NSW Office of Water should equip selected monitoring bores with water level-temperature-electrical conductivity loggers.

- The current surface and groundwater monitoring networks in the northern MDB is inadequate for the investigation and quantification of climate change impacts and it is recommended that network should be revamped to enable assessment of future climate change impacts on groundwater.
- The effect of data sampling interval and time discretisation on model outputs should be re-evaluated in the context of the Namoi Valley.
- Due to thermal retardation, the thermal method has limited application beyond 20 m from the bank of a river. For future investigations using heat as tracer, this major limitation should be taken into consideration. However, EC and temperature probes do have a larger zone of influence.
- Assessment of recharge and discharge process in this study was carried out at a reach scale. Further, research should be undertaken to quantify losses or gains at catchment and sub-catchment scales.

Chapter 7 Conclusions and Recommendations

"[Science] is not perfect. It can be misused. It is only a tool. But it is by far the best tool we have, self-correcting, ongoing, and applicable to everything. It has two rules. First: there are no sacred truths; all assumptions must be critically examined; arguments from authority are worthless. Second: whatever is inconsistent with the facts must be discarded or revised. ... The obvious is sometimes false; the unexpected is sometimes true."

Carl Sagan, 1985

After the millennium drought of 1997-2009, the issue of surface groundwater connectivity was an emerging water resources management priority within the MDB catchments. Consequently, as part of a PhD study, the thermal method was applied for the assessment of infiltration/exfiltration in selected catchments in NSW (Chapter 3). In the following paragraphs, the most important conclusions from Chapter 2 to Chapter 6 of this thesis are highlighted.

Although heat as an environmental tracer is usually applied overseas, its use in Australia has been limited so far to specific research sites, (Rau et al., 2014; Unland et al., 2013) and as part of this PhD study (Berhane et al., 2011). A summary of conclusions/recommendations based on field experimentation conducted in NSW are presented below.

7.1 Transmission Losses (TL)

In most climatic regions, evapotranspiration (ET) is the largest component of the water balance equation, accounting for an average of 70 percent of the consumption of annual precipitation in the United States, and up to 95 percent in arid and semi-arid climates, (Smith, 1998; Stafford-Smith and Morton, 1990; Stonestrom and Harrill, 2007) and Chapter 2 of this thesis.

Also in Australia, the inland lowland river systems are allogeneic. These rivers originate in the relatively watered upland areas, but drain for most of their river course in arid and semi-arid interior that yields little or no additional stream discharge. Due to high rate of evapotranspiration, infiltration and filling in of waterholes and billabongs, the inland river systems experience large-scale losses. This natural loss of water, which is an inherent characteristic of dryland river systems, is called transmission losses (TL). It is envisaged that TL may occur in any climate type, but it is most common in arid and semi-arid regions.

Transmission losses occur as part of the hydrological cycle and should not have a negative connotation as wastage of water. In fact, TL should be recognised as a natural process of the hydrological cycle, where water, energy and momentum move from one compartment to another. In addition, TL support ecosystems and recharge local aquifers and regional groundwater.

Available published reports for a different region, the Murray Darling Basin (MDB), suggest that TL can vary from 20 to 100% in this part of Australia. In fact, TL occurs not only in the inland river systems, but also on the coastal river systems, where river channels are predominantly intermittent or 'perennial'.

It was a challenging exercise to compile a literature review on TL for the MDB catchments, as the bulk of the literature reviewed appears biased toward humid-centric conceptual models and the inland river systems are considered to be groundwater dominated. However, based on the field observations in the Namoi Valley and literature reviews from previous studies, TL is considered to be an inherent characteristic of all catchments in NSW, irrespective of their geographic locations (coast or inland). After the introduction of water trading in

Australia, interest in TL has been revamped. Transmission losses along water supply routes are considered to be of four different types:

- Storage in waterholes/billabongs (ephemeral pools), channels and other wetlands;
- Evapotranspiration from the water surface;
- Seepage through the bottom of the river or channel;
- Leakage through the banks of the river or channel, and as overbank losses

After conducting a literature review on transmission losses and methods used for estimation infiltration, percolation and recharge, heat as an environmental tracer was found to be appropriate for this study (Chapter 2 of this thesis).

7.2 Recharge

Estimating of recharge is one of the prerequisite for water resources management and given the scarcity of hydrological information, it is a challenging exercise to estimate recharge at different time and spatial scales. Physical properties of basin, meteorological and hydrological factors influence groundwater recharge (Figure 7-1). However, the interplay between climate, landscape and hydrological factors are complex (Figure 7-1). Therefore, estimation of distributed recharge in arid and semi-arid environments is a challenging exercise (Chapter 2).

In the context of this research, the seepage rates along streambed were calculated using the thermal method for the designated sites in the Namoi Valley and Central West Catchments (Figure 4-1). Based on the hydrographs responses in the Namoi and the Northern MDB catchments, the magnitude of distributed recharge in productive agricultural areas, underlain by thick clay layers is considered to insignificant.

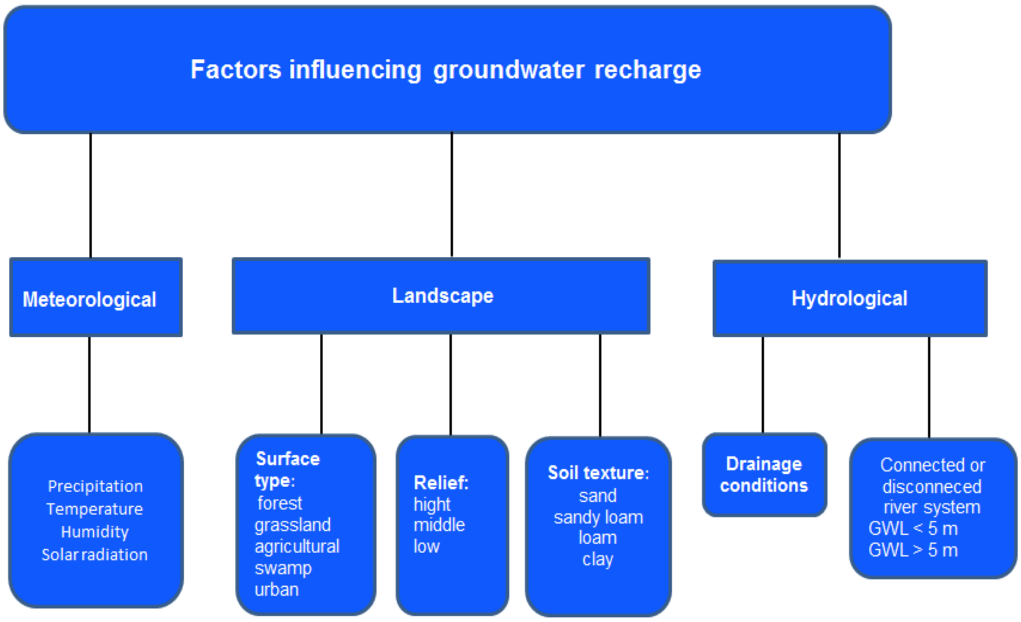


Figure 7-1. Factors influencing groundwater recharge (Modified after anonymous)

7.3 Ephemerality and Baseflow

Traditionally, a streamflow hydrograph consists of three components (surface runoff, interflow and baseflow). Surface runoff and interflow represent a quick components of a hydrograph. Quick flow originates from surface runoff while baseflow is sourced from local, intermediate and regional groundwater systems. The boundary between each of these sources of water is difficult to distinguish in practice. River channel precipitation and evapotranspiration also occur, but are generally small and indistinguishable from other components of streamflow (Sklash and Farvolden, 1979).

The inland river systems of the Northern MDB carry water only in direct response to effective precipitation. Horton overland flow dominates the stream hydrograph; contributions from subsurface flow do not exist or are insignificant (Dunne, 1983), Figure 7-2). The ephemeral nature of the creek systems, in combination with low creek/stream conductance, does not create a favourable environment for horizontal seepage. In addition, the creek does not intercept groundwater. Therefore, groundwater discharge from local, intermediate and regional groundwater systems are rare. From the author's experience, the only catchment where groundwater dominance does occur is along the Dumaresq River. However, seepages through preferential pathways or lateral inflows are not excluded. In addition, in connected systems, interflow is considered as part of the 'groundwater' component by some investigators. For instance, Marti et al. (2000), discusses in detail a conceptual model developed by for arid regions where the only input to the alluvial aquifer is through TL from the streambed. In this conceptual model, lateral water inputs are strictly from overland flow directly to the stream channel (Newman et al., 2006), cited the lack of dense vegetation and low permeability soils in the uplands as the reason for the dominance of overland flow over lateral subsurface flow. These models suggest that, for semi-arid drainages, it is critical to understand the importance, and model conceptualisation, of lateral inputs to the stream and aquifer and whether effluent or influent conditions prevail (Beven, 2002; Linsley et al., 1975).

On a lateral contribution to stream discharge, Beven (2002) further observed that overland flow processes do not easily explain extended recession limbs on discharge hydrographs from some semi-arid catchments. It is plausible that the shape could be as a result of subsurface runoff contributions, for example bank storage.

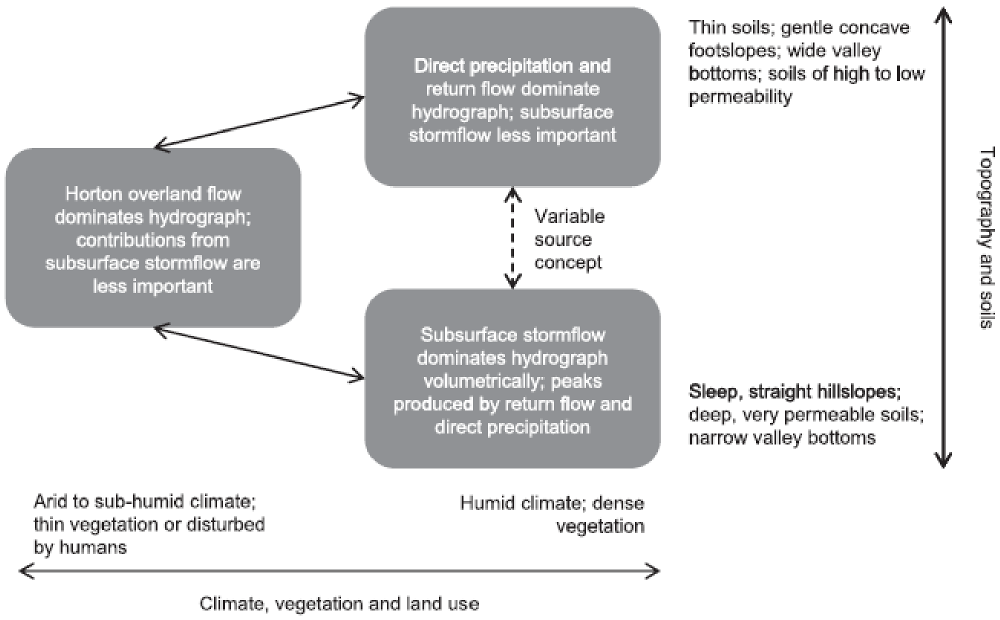


Figure 7-2. Dominant processes of runoff generation mechanisms at the hillslope scale (Dunne and Leopold, 1978; Dunne, 1983).

Based on the available hydro(geo)logical and geomorphological data sets, the Cockburn Valley does not show any groundwater dominance characteristics, at catchment and sub-catchment scales. For instance, the EC parameters measured at Kootingal and Mulla Gauging station vary in a wide range of values. This trend in EC may provide a clue on the lack of groundwater dominance.

Another major hydrogeological factor, which precludes groundwater dominance, is the lack of springs in the study area. Groundwater processes can strongly influence the drainage network. However, in the context of the study area there is no geomorphological information to suggest groundwater dominance on influencing river channel geomorphology.

Based on literature review and recent work carried out by the author in the Border Rivers, the Dumaresq River is probably the only catchment in the northern MDB which shows positive groundwater heads, along some reaches.

7.4 Heat as a hydrological tracer for assessment of infiltration

This PhD research work has successfully applied heat as an environmental tracer in different hydrogeomorphological conditions in NSW for assessment of infiltration and exfiltration. Based on the available hydr(geo)logical data sets, infiltration and ET are the dominant hydrological process in the Namoi Valley. Thus TL constitutes a large component of the water balance equation.

Prior to analysing the thermal data set from the experimental field sites in the Namoi and Central West catchments, an insight into hydrological processes of ephemeral and intermittent streams was developed by simulating hypothetical head distributions using the VS2DT package. The results from the VS2DT simulations suggest that the relationship between seepage and groundwater level is highly variable and non-linear for a disconnected river with a shallow water table setting. Similar to previous virtual simulations it was confirmed that the seepage flux reaches a constant value below a certain critical depth in disconnected river systems, for an isotropic and homogenous aquifer system. This critical depth of water table, where river seepage flux attains a constant value depends on channel geometric configuration and hydraulic conductivity of the streambed and underlying aquifers.

The main conclusions from this field based experimental work on the thermal method are:

- the method is easy to install and monitor. In addition, the sensors can easily be integrated into the existing monitoring network. Downward and upward fluxes can be estimated at a required temporal scale;
- the method is relatively in-expensive compared with chemical and isotopic methods;
- the method is in- sensitive to fluxes below 10^{-6} m/s (Figure 5.6). Flux values much less than 2×10^{-4} suggest that heat transport is dominated by conduction (Silliman et al., 1995). Conversely, values much greater than 2×10^{-4} suggest that heat transport is dominated by advection (Blasch et al., 2007). Based on the work carried out in the Cockburn valley, probably the threshold hydraulic conductivity is in the range of 10^{-5};
- from a communication point of view, experimentation results are clear to understand and are easily communicated with the stakeholders. Temperature is related easily to our daily activities, when compared to stable isotopes and other tracers used to estimate recharge and discharge processes.

Australia being the driest continent, variability of streamflow is a major factor in water resources and environmental management. In unregulated catchments, extended periods of cease to flow are a norm rather than exception for in land river systems of the MDB. For example, in the northern part of the MDB, no flow duration can extend for more than 370 days, which has a detrimental effect on the ecosystems (Marshall, 2006a). Even in the wettest part of Queensland, outside the MDB, rivers dry-up for an extended period, ranging from 20 to 180 days (Beven (2002); (Marshall, 2006a) . In NSW, a maximum no flow duration of 252 days occurred during the federation drought in 1902. In this type of environment, knowledge on TL is a prerequisite for effective water resources management.

As demonstrated in the Cockburn Valley, heat tracing technique can be used as a surrogate indicator to determine the status of connectivity within the streambed zone (5.4.2.1). The ability to understand when connectivity is lost and then re-established over a range of conditions will assist land and water managers to better manage water resources (Figure 5-15).

7.4.1 Inference of the Dynamic Nature of Streambed Conductance from a thermal streambed data

In the northern MDB region, summer runoff and flood events are prevalent when compared to winter runoff. During flood events, the probability of a de-clogging process increases and induces an increase in hydraulic conductivity. In tandem with an increase in hydraulic conductivity, streambed temperature increases in summer season, resulting in the increase of the hydraulic conductivity. For instance, stream temperature in the Cockburn River varied from 10 to 30°C in 2001, equivalent to a decrease in fluid viscosity of 40%. Thus, a variation in stream temperature influences infiltration rates at daily and seasonal scales. In the VS2DH package, the dependency of hydraulic conductivity with temperature due to viscosity changes is accounted for, using an empirical formula developed by (Kipp, 1987).

7.4.2 Is the issue of scale a puzzle in eco-hydrological sciences?

Previous studies on TL in the Namoi Valley were mainly qualitative or semi-qualitative in their approach and were based on a single method at sub-catchment scale or catchment scale. This may not necessarily provide an insight into the complex hydrological processes that take place at more local scales (reach and streambed). Exchanges of surface water and groundwater also occur at the channel bed scale. That is, local shallow surface water circulations into the underlying sediments create areas of groundwater infiltration and exfiltration within zones generally characterised as gaining and losing stream sections (Woessner, 2000).

Due to geological and geomorphological factors, the hydraulic characteristics of a groundwater system show high variability in space. At a point scale, hydrogeological data set may be available to characterise aquifer parameters. However, these data sets are rarely available at sub-catchment and catchment scales. In any case, together with spatial variability of the aquifer physical parameters, boundary conditions are additional sources of unknown in developing catchment scale models.

In the case of the Namoi Valley, hydrogeological cross-sections at catchment and sub-catchment scales are presented in Chapter 2 (Figure 2-19) and Chapter 6 (Figure 6-27). The hydrogeological are similar to hydrogeological cross-sections reported from the semi-arid region of the western US (Stewart-Deaker et al., 2008) and the Nile region of Africa. All of these cross-sections do not show a simple relationship between topography and water table or the water table is not a replica of topography. Thus, the requirements for the occurrence of a nested groundwater flow as postulated by (Toth, 1963) are not met. More interestingly, without a measurable discharge (baseflow) component, is it possible to constrain regional groundwater flow models in arid and semi-arid regions? Based on the recent work carried out in the Northern part of Queensland, sub-marine groundwater discharge is considered one of the modes of groundwater discharge (Leach, 2015).

Monthly stress periods are usually used to mimic hydrological process at catchment and sub-catchment scales in humid environments. However, for estimation of recharge/infiltration rates of ephemeral river systems, a monthly stress period may not necessarily capture episodic runoff events. Consequently, the effect of temporal domain discretization on infiltration rates was evaluated using an hourly and daily time steps for the Kootingal transect. Within the streambed zone, above the critical water table depth (see Chapter 3), vertical fluxes simulated using an hourly stress period were greater, when compared to a daily time discretization.

Last but not least, it is true that conjunctive water resources management is the desired option. However, from the author's perspective, management scenarios should not blur the distinction between surface and groundwater resources. The non-dominance of groundwater should be explored in the context of four major areas:

- groundwater hydraulics (groundwater heads VS river stages), including occurrence of springs. More importantly, hydrogeological characteristics should be considered as a foundation in studying surface/groundwater connectivity;
- the quality of groundwater and surface waters;
- the instream ecology;
- the geomorphology of sediments.

Nevertheless, from management perspective surface groundwater connectivity should be studied at a regional scale. It is only at a regional scale that social, economic and ecological impacts can be assessed (Barthel, 2014).

7.4.3 Non-linear hydrologic processes

The Kootingal gauging station was established in 2006, for assessment of environmental flow in the Cockburn Valley; given the availability of stream stage data (2006 to 2015) and calculated vertical seepages (2008-2012, Chapter 4) for this site; an attempt was made to develop a relationship between these two parameters. Due to highly non-linear characteristics of the unsaturated zone, a cross plot of vertical seepage against stream stage exhibits hysteresis (Figure 7-3). Thus no attempt has been made to develop empirical relationship between stream stage and vertical seepage.

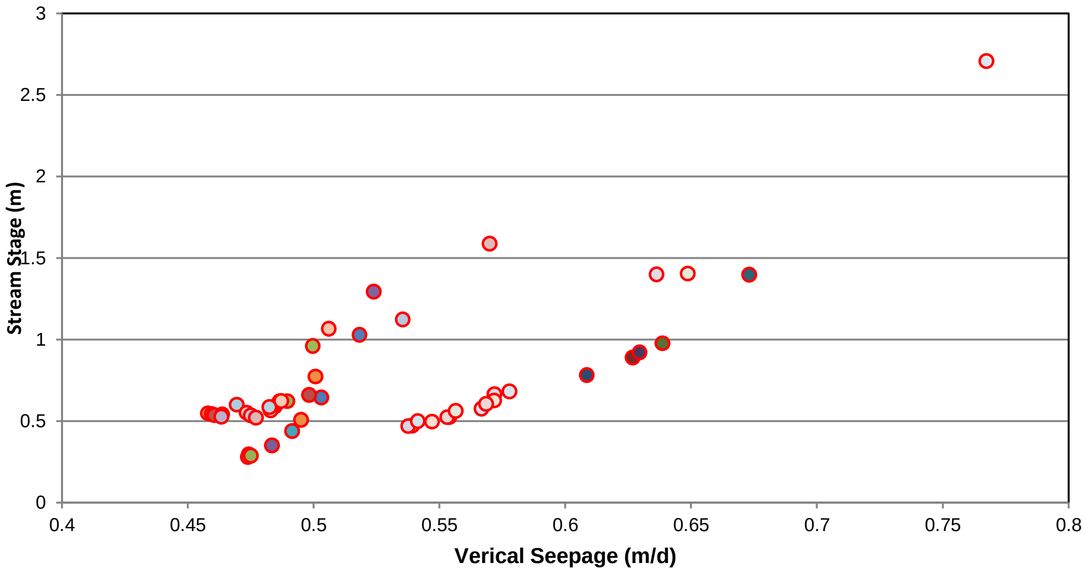


Figure 7-3. A non-linear relationship between vertical seepage and vertical seepage: Kootingal Transect

Unsteady groundwater flow equation becomes non-linear with serious difficulties in its solution (Kraijenhoff van de Leur, 1958, Dooge, 2005).

7.4.4 Issue of hydro(geo)logical complexity

Toth (2006) in his recent book stated: "In nature, the various environmental parameters can combine in a virtually unlimited number of ways, each of them modifying a groundwater – induced basic process or phenomenon in a different way."

Complexity theory is applied to study the behaviour of systems in physics, biology the social sciences and economics (Manson, 2001; Johnson, 2007; Bloomfield 2008). It is appropriate to apply mathematical equations to solve hydrological process. However, to assume differential equations mimic all processes (physical, biological and chemical) is an exaggeration and does not make sense. For instances, in Chapter 4 of this thesis, I have highlighted a simple positive feedback mechanism, which occurs frequently in hydrology. Other complexity issues addressed in the thesis are succinctly shown in Table 7-1.

Table 7-1. Complexity of groundwater systems as captured/highlighted in this thesis

Feature	Description	Chapter(s)
Semi-closed or closed GW systems	In arid and semi-arid regions, modern recharge is insignificant (< 5 mm) and in some environments the GW compartment should be considered as closed or semi closed systems.	Chapter II:
Unsaturated zone	In humid environments, the role of unsaturated zone is less important than in arid and semi-arid environments. In contrast, for groundwater management purposes, the unsaturated zone cannot be ignored. However, this mathematically challenging exercise.	Chapter III, IV and V of the thesis
Baseflow	Groundwater systems in arid and semi-arid regions are not groundwater dominated. However, under favourable climatic conditions, the groundwater dynamics can switch from losing to gaining conditions (For example, Lower Peel Valley).	Chapter III and V
Feedback	A feedback loop is a mechanism by which change in a variable results in either dampening (negative feedback) or amplification (positive) of the change of state. https://serc.carleton.edu/NAGTWorkshops/complexsystems/introduction.html#feedback	De-clogging process, which takes place during flood events, is an example of positive feedback, where removal of fine sediments increases permeability of streambed sediments. Negative feedbacks are associated with sediment clogging during low flow events (Chapter IV).
Nonlinear behaviour and relationships	Complex nonlinear systems are inherently unpredictable a physical system may exhibit a deterministic chaotic behaviour (Faybishenko, 2004). In some cases, small perturbations in the system may cause large effects, a proportional effect or no effect at all (Phillips, 2006).	Groundwater systems demonstrate a wide range of nonlinear characteristics (Chapter III, IV and V).
Issue of scale	The issue of scale is very important in any surface and groundwater connectivity study. Local effects may differ from reach behaviour in both space and time.	Chapter V

7.5 Recommendations

The final recommendations for this study are:

- In all climatic regions, groundwater recharge is one of the most difficult parameters to estimate. At a catchment and sub-catchment scale, infiltration/recharge rates should be estimated using multiple methods (Chapter 2 of this thesis).
- As a recent study in the Border Rivers suggest, at a basin scale, the Gravity Recovery and Climate Experiment (GRACE) satellites was applied to monitor changes in total water storage (TWS) in saturated and unsaturated zones.
- Hydrologic processes in disconnected groundwater systems of water-limited environments unfold over longer timescales than those in surface and near-surface soils. Quantification of recharge is thus beyond simplistic models and would require deep vertical profiles of water content and water potential (to ascertain if gradients favoured upward versus downward water movement) along with chloride profiles (to quantify recharge by the chloride mass balance method, In order to constrain recharge estimates, multiple methods (physical, chemical and isotopic) should be used;
- The magnitude of infiltration at hourly and daily time scales was shown to depend on the depth to the water table Figure 6-39. Thicker unsaturated zones overlain by thick aquitards are associated with lag-times that have potential to record many years of moisture and salt profiles;
- A threshold hydraulic conductivity where the pattern of pressure head distributions changes from a bell shaped to a non-bell shaped distribution takes place when K approaches 10-8 m/s. The velocity vector is oriented predominantly vertical in the cross-section;

- There is inconsistency in literature regarding the definition of a clogging layer and this should be rectified. Is a clogging layer a physical factor or a fudge parameter used in modelling? In addition, how does water move in a thick aquitard layer overlying most productive aquifers in the Northern MDB catchments?

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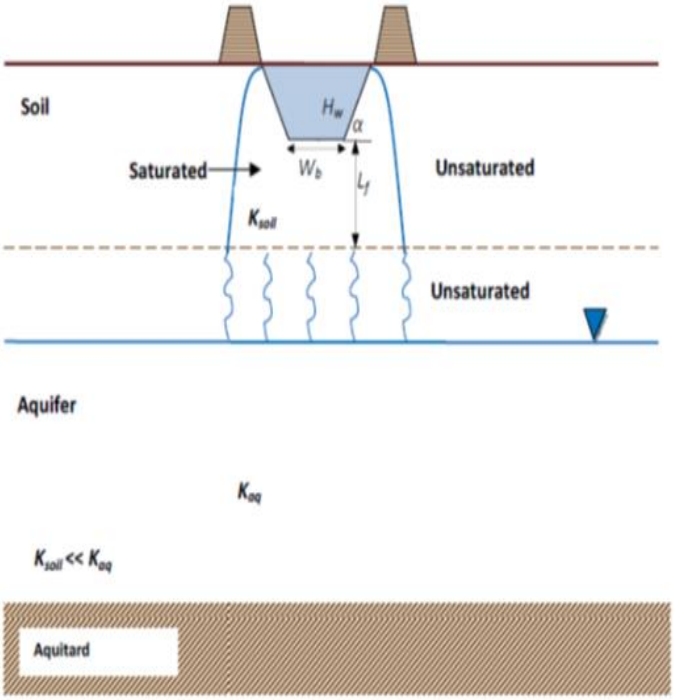
Appendix A: Drillers' Logs for the Kootingal Transect

WORK_NO	FROM_DEPTH	TO_DEPTH	DRILLERS_DESCRIPTION	ROCK_TYPE
GW093036	0	3	Black Clay	Black Bands
GW093036	3	8	Brown Clay and Gravel	Brown Clay Bands
GW093036	8	11	Fine Gravel	Gravel
GW093036	11	12	Shale Black	Shale
GW093037	0	2	Brown Clay	Brown Clay Bands
GW093037	2	10	Brown Clay bound Gravel	Brown Clay Bands
GW093037	10	19	Gravel	Gravel
GW093037	19	21	Spotted Gravel	Gravel
GW093038	0	6	Brown Clay/Gravel	Brown Clay Bands
GW093038	6	13	Sandy Gravel	Sandy
GW093038	13	16	Clay Sand and Gravel	Clayey Sand
GW093038	16	17	Spotted Granite	Granite
GW093039	0	11	Brown Clay	Brown Clay Bands
GW093039	11	12	Sandy Clay	Sandy Clay
GW093039	12	15.5	River Gravel	Gravel
GW093039	15.5	16	Shale Yellow	Shale
GW093040	0	1	Sandy Top Soil	Topsoil
GW093040	1	12	Brown Clay	Brown Clay Bands
GW093040	12	13	Decomposed Granite and Clay	Decomposed
GW093040	13	16	River Gravel	Gravel
GW093040	16	17	Moombi Granite	Granite

Appendix B: Seepage Estimation using Bouwer's Analytic Method

The seepage rate for a disconnected river system can be calculated by applying Darcy's Law to the soil layer and considering the negative pressure head at the base of the soil, as described by Bouwer (2002b):



Parameters			Units
Wetted perimeter of the channel	Wp	25	m
Vertical Hydraulic conductivity	Ksoil	0.01	m/d
Height of the water in the channel	Hw	2	m
Thickness of the streambed	Lf	0.5	m
Negative pressure head	Hwe	-0.35	m

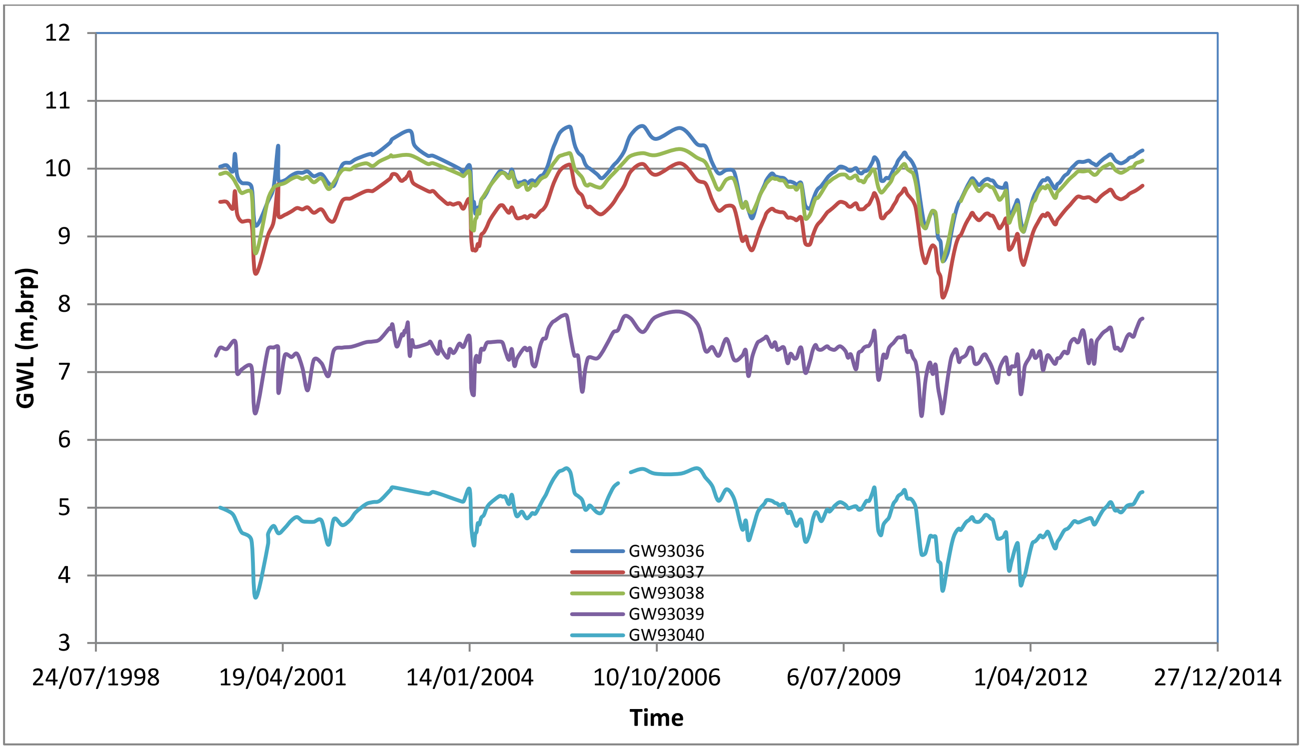
Stream seepage can be calculated using Bouwer's equation of the form:

(KL/d/m)

$$q = W_p K_{soil} \frac{(H_w + L_f - H_{we})}{L_f}$$

q 1.425

Appendix C: Groundwater Level Trends: Kootingal Transect



Appendix D: Low Flow Analysis: Cockburn Valley (9-10 October 2006)